S100 22 and 23

THE OPEN UNIVERSITY

Science Foundation Course Units 22 and 23

The Earth, Its Shape, Internal Structure and Composition / The Earth's Magnetic Field

The Open University

Science Foundation Course Unit 22

THE EARTH: ITS SHAPE, INTERNAL STRUCTURE AND COMPOSITION

Prepared by the Science Foundation Course Team

THE OPEN UNIVERSITY PRESS

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Contents

General Introduction to Units 22 to 27

As you pick up the text for this week, you may well be thinking: now we've finished the biology, here come the Earth scientists with their bit of the Course. And you would be right—up to a point.

There are, nevertheless, some good reasons for switching attention to the Earth just at this point in our story. Think back for a moment about the ground we have covered over the past twenty-one Units. We set the scene with a discussion of the origins and methods of science, and then we examined the material environment, starting with some fundamental properties of matter in motion, then going on, up the scale of complexity, from atoms, to molecules, to cells, to organisms, to populations. And finally, in Unit 21, we remarked upon the biochemical unity within the vast diversity of living matter, with the inference of a common origin in a 'primordial soup'. The soup in question was on a particular planet—our planet, the Earth. It seems logical to stop and ask at this point: how did the soup get there? What were the special features of this planet that made this most extraordinary thing we call 'life' possible? If the story of the evolution of life started in the soup, where did the story of the evolution of the soup start? This line of questioning leads naturally into a study of the Earth, its structure, its history and its origins.

Could we, perhaps, start by looking for the remains of past forms of life? Traces of life forms have been found in rocks older than 3 000 million years (3 000 Ma). There may be organic compounds in meteorite material that is 4 500 Ma old. So the 'pre-history' of life may have begun very early in the history of the Earth itself. Its traces can be found, not in 'ordinary' fossils but in what might be called 'biochemical fossils', for, as you have seen, organisms would have been preceded by organic-compound synthesis of some kind.

But, as well as the tangible record of organic activity left behind in the rocks of the Earth, we need to know about the past conditions or environments on the Earth's surface. Then perhaps we may begin to understand when and how life originated, and how it subsequently evolved. We now propose to study the Earth to try and find out something of its long history. One, but only one, of our aims will be to try and explain the origins of life. Other features of our planet's surface, such as why continents are where they are, why earthquakes occur, why volcanoes erupt, why there are hurricanes, why deposits of useful minerals such as coal, oil and metallic ores etc. occur in special places, can all be understood more fully when we have a knowledge of the Earth's history. Indeed, our knowledge is becoming so precise that we can often predict when and where natural hazards such as volcanic eruptions and hurricanes will occur and where economic mineral deposits are most likely to be found.

So now we are going to focus our attention on the planet we inhabit—the Earth. It is literally under our noses, yet until very recently we had less idea of its internal structure than of the motions of the other planets in our Solar System. Why is this so? The clue lies in the use of the terms geology and Earth sciences.

Have you wondered why, in this Course, we prefer 'Earth sciences' to 'geology'?

Geology, although literally meaning the study of the Earth, has by common usage become the study of the rocks of the visible or accessible parts of the Earth's crust. Present-day investigation of the Earth is much more than

this. For instance, the sophisticated methods of geophysics and geochemistry are used to determine the physical properties of the materials deep within the Earth and the chemical processes active there. Knowledge of the Earth as a planet is being advanced as much by the techniques of physics and chemistry as by the traditional observational methods of geology. For the last 300 years geologists, although speculating about the Earth's interior, have been primarily concerned with its crust. It is only since the turn of the century, and particularly since the Second World War, that the advanced techniques of geophysics and geochemistry have been developed. Geology, geophysics and geochemistry are, collectively, the Earth sciences. It is through the combination of these sciences that the many fascinating problems of the Earth's structure and history will be solved.

The next few Units may help you to appreciate the particular difficulties and complexities that confront the Earth scientist, and the techniques and ideas he has developed to try to overcome them. To start with, there are three questions we will attempt to answer in Units 22 and 23:

1 What is the shape of the Earth and why does it have such a shape?
2 What is its internal structure and composition?
3 Why does the Earth appear to act as a magnet?

Before we get to grips with these questions, there is a point you should keep in mind: we are going to be concerned primarily with the physical properties of the Earth *as it is today*. However, it will become more and more evident as our study progresses that the Earth is *not* a static, unchanging lump of matter, but a dynamic body that is changing now and has been changing ever since it was formed.

Study Comment

Unit 22 is somewhat longer than the average length of the Units you have studied so far. We realize, therefore, that it may take you more than a week to get through it. However, Unit 23 is deliberately shorter than average; so at the end of the two-week period you should just about be level again. Do not worry, then, if you find Unit 22 taking longer to study than you had expected.

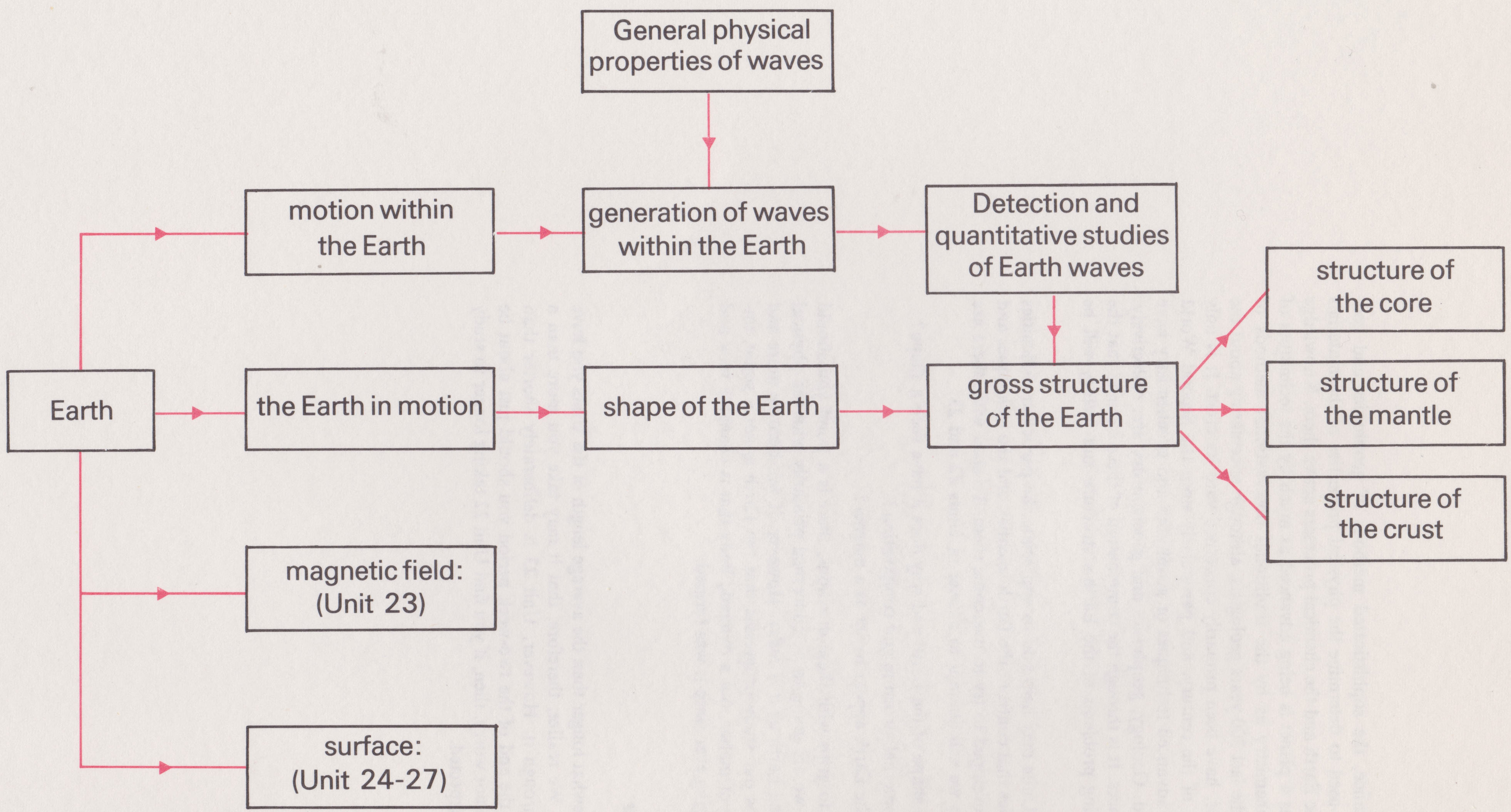

General physical properties of waves
Earth
motion within the Earth
the Earth in motion
magnetic field: (Unit 23)
surface: (Unit 24-27)
generation of waves within the Earth
shape of the Earth
Detection and quantitative studies of Earth waves
gross structure of the Earth
structure of the core
structure of the mantle
structure of the crust

Table A

List of Scientific Terms, Concepts and Principles used in Unit 22

Taken as pre-requisites			Introduced in this Unit			
1	2		3		4	
Assumed from general knowledge	**Introduced in a previous Unit**	**Unit No.**	**Developed in this Unit**	**Page No.**	**Developed in a later Unit**	**Unit No.**
meteorite	diffraction	2	Titius-Bode Law	11		
fossils	angular velocity	3	asteroids	12		
planet	newton	3	oblate spheroid	12		
solar system	law of gravitation	4	precession	16		
magnet	biochemical	21	Chandler wobble	17		
satellites			nutation	17		
latitude			seismology	18		
			earthquake waves	19		
			seismometers	19		
			elastic limit	20		
			Hooke's Law	20		
			perfectly elastic	20		
			perfectly plastic	20		
			strain	20		
			yield point	20		
			stress	21		
			Young's modulus	21		
			axial modulus	22		
			Poisson's ratio	22		
			rigidity modulus	23		
			shear strain	23		
			longitudinal (P) wave	24		
			shear (S) wave	24		
			earthquake focus	26		
			earthquake zones	26		
			aftershocks	27		
			earthquake intensity	27		
			earthquake magnitude	27		
			foreshocks	27		
			isoseismal line	27		
			principal shock	27		
			seismogram	28		
			wave propagation	29		
			interference	30		
			superposition principle	30		
			Huygens' principle	31		
			angle of incidence	32		
			angle of reflection	32		
			laws of reflection and refraction	32		
			secondary wavelets	32		
			angle of refraction	34		
			refractive index	34		
			Snell's Law	34		
			critical angle	35		
			total internal reflection	35		
			epicentral angle	36		
			body waves	37		
			core-mantle boundary	42		
			discontinuity	42		
			inner core	43		
			Mohole	45		
			Mohorovičić discontinuity	45	*(continued)*	

Taken as pre-requisites			Introduced in this Unit			
1	2		3		4	
Assumed from general knowledge	Introduced in a previous Unit	Unit No.	Developed in this Unit	Page No.	Developed in a later Unit	Unit No.
			Conrad discontinuity	46		
			sial (sialic layer)	46		
			phase change	47		
			sima (simatic layer)	47		
			low-velocity layer	48		
			incipient melting-point curve	49		
			asthenosphere	50		
			lithosphere	50		
			mesosphere	50		

Any scientific terms used in this Unit but not listed above are marked thus † and defined in the glossary (p. 61).

Objectives

When you have finished your work for this Unit, you should be able to:

1 Define correctly, or recognize the best definitions of, or distinguish between true and false statements concerning all the terms, concepts and principles in column 3 of Table A.

2 Given data on the Earth's overall density, recognize true and false statements concerning the relative densities of the Earth's crust and interior.

3 Describe the kind of data concerning the Earth's interior that can be deduced from the study of earthquakes.

4 Given the necessary data, perform simple calculations using Hooke's law, Huygens' principle, the laws of reflection and refraction, and Snell's law.

5 Draw and label correctly a cross-sectional diagram of the Earth through its centre.

6 Illustrate diagrammatically the differences between P and S waves and compare and contrast their properties,

7 Make deductions concerning travel-time data, the Earth's density, the speed of P and S waves, and the presence of discontinuities from given depth-velocity curves.

8 Apply information in the Unit to given problems concerning models of the Earth's structure.

Table 1 The Solar System

Planet	Mean distance from Sun in astronomical units (See Note 1)	Approximate diameter (Earth = 1)	Mass (Earth = 1)	Average density (10^3 kg m^{-3})	No. of satellites known (See Note 2)	Orbital period (years)	Inclination of orbit to ecliptic† (degrees)	Inclination of equator to orbit (degrees)	Rotation rate (See Note 3)
Mercury	0.38	0.4	0.05	5.1	0	0.24	7.0	?	59 days
Venus	0.72	1.0	0.9	5.3	0	0.61	3.4	23	243 days
Earth	1.00	1.0	1.0	5.52	1	1.00	0.0	23	1 day
Mars	1.52	0.5	0.11	3.94	2	1.88	1.9	24	24.5 hours
Jupiter	5.20	11	318	1.33	12	11.86	1.3	3	10 hours
Saturn	9.54	9.54	95	0.69	10	29.46	2.5	27	10 hours
Uranus	19.18	4	15	1.56	5	84.01	0.8	98	11 hours
Neptune	30.06	4	17	2.27	2	164.79	1.8	29	16 hours
Pluto	30.44	?0.5	?0.1	?	0	247.69	17.2	?	6 days

Notes:

(1) 1 astronomical unit = 1.496×10^{11} m (93×10^6 miles approx.).

(2) Nearly all satellites move in their orbits in the same direction as the rotation of their accompanying planet, but the outermost four planets of Jupiter move in the opposite direction. Most of the satellites have orbits very close to the equatorial plane of their planet; the main exception is the Moon, whose orbit is nearly coincident with the plane of the *ecliptic*†.

(3) The direction of rotation of the planets and their orbital directions coincide, except for Venus.

Section 1

22.1 The Earth

In the general introduction to this and the next five Units, we said that we would be concerned with the Earth and its history. In this Unit, we shall start by noting quite briefly and simply *where* the Earth is with respect to the rest of the Solar System, the Earth's immediate environment.

Then we shall ask a series of questions:

(i) What is the size and shape of the Earth?

(ii) What is its mass and mean density?

(iii) How does the Earth move?

(iv) How does one find out anything about the internal structure of the Earth?

At this point we shall stop to think about waves and how they are propagated in various substances, especially about how the velocity of propagation depends upon the rigidity, compressibility and density of the material through which they pass. For armed with these simple physical concepts, and with some sufficiently sensitive and accurate instruments, the geophysicist has in recent years assembled a quite detailed and convincing picture of the internal structure of the Earth—its division by major discontinuities into a core, a mantle and a crust, and a great deal about their composition and structure.

In the concluding sections of this Unit we shall follow through the process of assembly of this 'model' of the Earth's interior. When you come to these sections, take careful note of the underlying method—of starting with a simple 'model' and refining it in the light of experimental evidence.

Section 2

22.2 The Earth in the Solar System

Before we get down to our study of the Earth itself, we should at least take a glance at its context—its place in the Solar System. Table 1 summarizes some of the most important data about the planets that comprise this System; and from it you will see that the Earth forms a very small and insignificant part. (You do not, incidentally, have to memorize the figures in Table 1.)

In 1766 the astronomer Titius noticed that the distance* of the planets from the Sun ran as a progression expressed as

$$\frac{4+0}{10}, \frac{4+3}{10}, \frac{4+6}{10}, \frac{4+12}{10}, \frac{4+24}{10}, \frac{4+48}{10}, \ldots\ldots$$

that is,

0.4, 0.7, 1.0, 1.6, 2.8, 5.2, ...

except that there is a gap in the progression between Mars and Jupiter. This discovery was made better known by another German, Bode, some six years later, and so is often referred to as the *Titius-Bode 'Law'*. There is no apparent explanation for the 'law', which really is a 'rule of the thumb' and does not work too well for the outermost planets.

Titius-Bode Law

* *These distances are measured in* astronomical units. *An astronomical unit is the mean distance from the centre of the Earth to the centre of the Sun. See Note* 1 *of Table* 1.

If you examine the first column of Table 1 you can see that this 'law' suggests that there should be another planet between Mars and Jupiter. The prediction made by this 'law' was apparently confirmed in 1801 when a small planet named Ceres was discovered about 2.8 astronomical units from the Sun. Further investigations revealed the surprising result that many more small planetary bodies existed at this distance from the Sun.

Thus the gap in the Titius-Bode 'law' was filled not with one planet but with many. These planets are now known as the minor planets or the *asteroids*. Today it is estimated that there are over 40 000 minor planets that could be observed by a 100-inch telescope. They range in size up to a maximum of 785 km diameter in the case of Ceres.

asteroids

Accurate observations of tracks formed by meteorites* entering the Earth's atmosphere have shown that the orbits of these bodies reach out to the asteroid belt, so it is sometimes assumed that this is where they come from. It has been suggested that the asteroids are the result of the break up of a planet, or perhaps several planets. If this is so, it means that the study of meteoritic material recovered from the Earth's surface may offer clues to the internal composition of other planets, among them the Earth, as we shall see later in the Course.

From the data given for the diameters, masses and average densities can you see any distinct grouping of the planets?

The planets seem to be grouped into two families, the inner, or terrestrial, planets and the outer planets. The terrestrial planets, Mercury, Venus, Earth and Mars, are all fairly small, having diameters the same or less than that of the Earth, and are generally rather dense bodies, being four or more times as dense as some of the outer planets.

Section 3

22.3 The Earth as a Body

22.3.1 Is the Earth spherical?

Millions of people have seen photographs of the Earth taken from satellites, from space vehicles or from the Moon; it is evidently spherical. So is the Moon, and judging from photographs taken through telescopes, or, as in the case of the planet Mars, from space vehicles like those of the 'Mariner' series, all the planets are more or less spherical. But is the Earth *exactly* spherical?

Let us see what relevant information we have. Accurate measurements have shown that the length of one degree of latitude is less near the equator than at the poles by about one part in a hundred. This, in turn, means that the sphere is more strongly curved at the equator than in the polar regions. In other words, the Earth is not a perfect sphere at all but has a shape known as an *oblate spheroid*, a sphere flattened at the poles (Fig. 1).

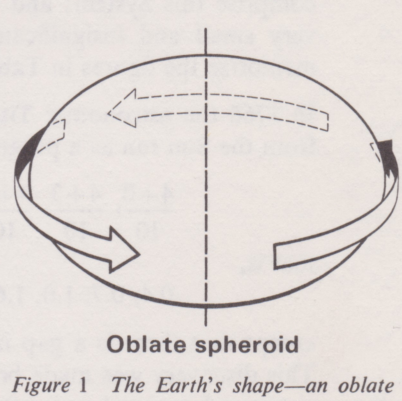


Figure 1 The Earth's shape—an oblate spheroid, obtained by rotating an ellipse about its minor (that is, shorter) axis.

The shape of the Earth can be determined by purely astronomical methods, such as an analysis of the irregularity of the motion of the Moon. A more

* *For further details of meteorites see the article by Dr. B. Mason in* Understanding the Earth (*Chapter* 8).

recent and more accurate method has been to study the trajectories of artificial satellites orbiting around the Earth. Both methods depend upon the gravitational interaction between the Earth and the satellite, natural or artificial. The force on the satellite, and hence its motion around the Earth, is not quite the same as it would be if the Earth were exactly spherical.

We can define the oblateness, f, of the Earth's spheroid as:

$$f = \frac{(r_e - r_p)}{r_e}$$

where r_e is the equatorial radius and r_p the polar radius.

The most recently reported value for f, obtained from satellite data, is

$$f = 1/(298.25 \pm 0.03) = 0.003\,352\,9 \pm 0.000\,000\,3$$

The actual values of the equatorial and polar radii are:

$$r_e = 6378.16 \pm 0.15 \text{ km}$$
$$r_p = 6356.77 \pm 0.15 \text{ km}$$

The Earth's ellipsoid is shown, schematically, in Figure 2.

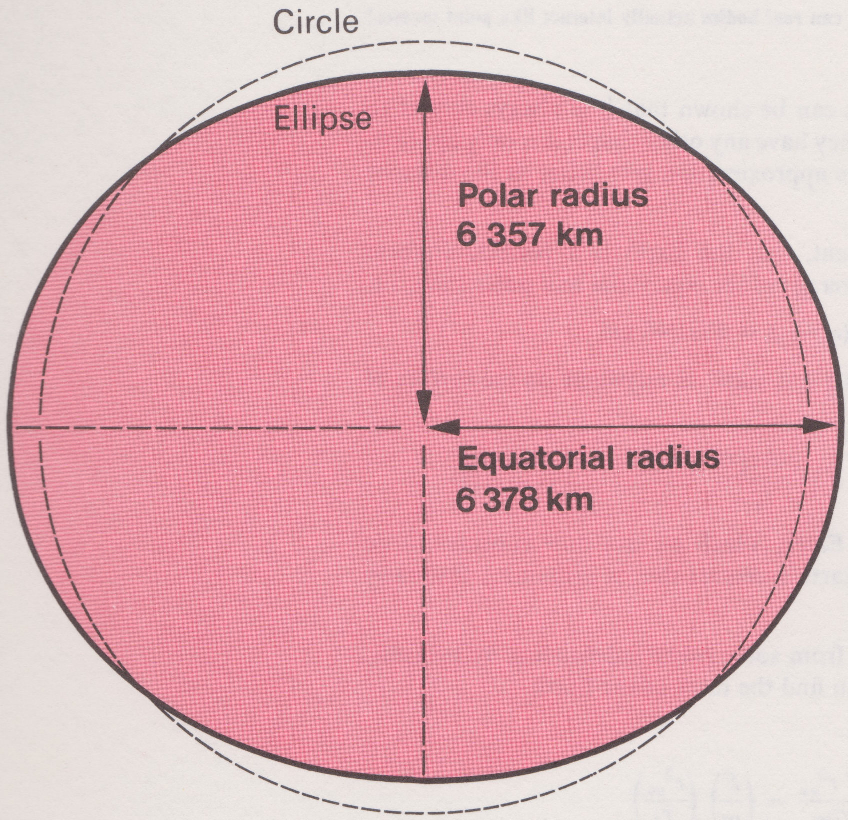


Figure 2 The Earth's ellipsoid.

The ellipticity of the Earth is greatly exaggerated in Figure 2. As you can verify for yourself, if the Earth were a globe 45 cm in diameter, then the difference between its equatorial and polar diameters would be a mere 1.6 mm, and all the topographic irregularities, such as Mount Everest, would be obliterated by a coat of paint!

22.3.2 The mass of the Earth

You may remember from Unit 4 that the gravitational force of attraction, F, between two 'point masses', m_1 and m_2, separated by a distance r, is:

$$F = \frac{Gm_1m_2}{r^2}$$

where the gravitational constant G has the value:

$$G = 6.668 \times 10^{-11} \text{ N m}^2 \text{ kg}^{-2}$$

if the masses m_1 and m_2 are in kg and r is in metres.

Could we somehow use this relation to determine the mass of the Earth? Could we, for instance, arrange to measure the force on some test mass, m, at a distance r away from the Earth, and hence find the mass of the Earth from the above formula? This inevitably raises the question: can we regard the Earth as a point mass?

Do you remember what is meant by a point mass?

If we imagine all the mass of a body, such as the Earth, to be concentrated at a point (which we call its *centre of mass*, or *centre of gravity*), and the mass of another body, such as the Moon, to be similarly concentrated at another point, then if the distance between these two points is r and the masses are m_1 and m_2, the force will be given by the law of gravitation.

Under what circumstances can *real* bodies actually interact like point masses?

If they are uniform spheres, it can be shown that it is always correct to treat them as point masses. If they have any other shape, it is only approximately correct to do so, but the approximation gets better as the distance between the bodies increases.

Let us pretend, for the moment, that the Earth is a perfect, uniform sphere, of radius equal to the average of the equatorial and polar radii, i.e.

$$r_{av} = 1/2\,(r_e + r_p) = 6367.46 \text{ km}$$

Then the gravitational force on a test mass, m, anywhere on the surface of the Earth would be:

$$F = \frac{Gm_E m}{r_{av}^2} \quad \ldots\ldots\ldots\ldots\ldots\ldots(1)$$

where m_E is the mass of the Earth, which we can now consider to be concentrated at a point (the Earth's centre) that is distant r_{av} from any point on the surface.

So if we know the value of G from some other independent experiment, and if we know r_{av}, then we can find the mass of the Earth.

Rearranging equation (1):

$$m_E = \frac{F\,r^2_{av}}{Gm} = \left(\frac{F}{m}\right)\left(\frac{r^2_{av}}{G}\right)$$

The mass of the Earth is then obtained by measuring the gravitational force F on the test mass m, or in other words, the gravitational acceleration

$$g = \frac{F}{m}$$

Note that in this imaginary experiment to 'weigh the Earth', our measuring instrument measures only the *gravitational* force.

Can you think of any other force that might act on a test mass on the Earth's surface?
What about the effect of the Earth's rotation?

If you have forgotten about centripetal force, refer back to Unit 3.

The effect of the Earth's rotation would be to reduce the apparent weight, or the force on a test mass at the equator, but not at the poles. At the equator, the centripetal force would be:

$$F_c = m\, r_{av}\, \omega^2$$

where ω is the angular velocity of rotation of the Earth.

(We use r_{av}, rather than the equatorial radius r_e, to remind you that we are still discussing a hypothetical spherical Earth.)

So at the equator, we would not measure:

$$g = \frac{F}{m} = \frac{Gm_E}{r_{av}^2}$$

but instead:

$$g' = \frac{F}{m} - \frac{F_c}{m} = \frac{Gm_E}{r_{av}^2} - r_{av}\omega^2$$

But at the poles, where there will be no centripetal force, we would still measure:

$$g = \frac{Gm_E}{r_{av}^2}$$

Why is there no centripetal force at the poles? Is it because the angular velocity is zero there?

No! The angular velocity must be the same everywhere, otherwise the Earth would be in the process of screwing itself up into a corkscrew! It is, of course, because the radius of the circle of rotation becomes zero at the poles, i.e. on the axis of rotation.

How big is the centripetal acceleration compared with the gravitational acceleration?

We know that the gravitational acceleration varies from place to place on the Earth's surface, but its average value is about 9.806 $m\ s^{-2}$. So the centripetal acceleration is about 0.3 per cent of the gravitational acceleration. Thus, if we wanted to measure the Earth's mass in this way, to a precision of better than 0.3 per cent, we would have to do our measurement at the pole, or else take the centripetal acceleration into account. (Some further consequences of the Earth's rotation are discussed in Appendix 1 (White).)

The Earth rotates once about its axis in approximately 24 hours—actually in 23 hours 56 minutes 4 seconds, that is, 86 164 seconds.

So $\omega = 2\pi/86\ 164 = 7.292\ 12 \times 10^{-5}$ rad s^{-1}

Hence $r_{av}\, \omega^2 = 6.367\ 46 \times 10^6 \times (7.292\ 12 \times 10^{-5})^2$

$= 0.031\ 9\ m\ s^{-2}$

So far, we have been pretending that the Earth is a perfect sphere. We have seen that it should be possible, in principle, to measure the mass of such an Earth by measuring the acceleration of a test mass at the pole—or elsewhere, if we make the proper correction for the centripetal acceleration. But the real Earth is *not* a perfect sphere, and, as we mentioned earlier, it is not strictly correct to use the law of gravitation in such a simple way in this case, because the Earth cannot be treated simply as a point mass, m_E, concentrated at its centre.

To measure the mass of the real, spheroidal Earth by measuring the acceleration of a test mass at various points on its surface, one has to modify the simple formula to take into account the fact that the Earth is not spherical. The detailed calculations are fairly complicated and we need not be concerned with them. The basic principle is, however, just the same as the one we have discussed.

The latest value for the product (Gm_E), determined in this way, is

$$Gm_E = 3.986\,026 \times 10^{14}\ m^3\ s^{-2}$$

You can get an idea of the error that would be introduced by using the simple law of gravitation to find m_E, if you take the measured value of the acceleration at the poles (where you do not have to worry about centripetal effects) which is:

$$g_p = 9.832\,166\ m\ s^{-2}$$

and compare it with what you would deduce from the law of gravitation:

$$g_p = \frac{F_p}{m} = \frac{Gm_E}{r_p^2} = \frac{3.986\,026 \times 10^{14}}{(6.356\,77 \times 10^6)^2} = 9.864\,31\ m\ s^{-2}$$

So the error is:

$$\frac{(9.864\,31 - 9.832\,17)}{9.832\,17} \times 100 = 0.327 \text{ per cent.}$$

This is about the same as the percentage oblateness of the Earth's spheroid, $100f = 0.335$ per cent.

In section 22.3.1, we mentioned that the most accurate way of determining the shape of the Earth is by studying the motion of artificial satellites in orbit around the Earth. Since a satellite is not attached to the surface of the Earth, although it is 'attached' to the Earth's centre of gravity, it is not affected by the Earth's rotation. But variations in the gravitational attraction of the Earth, due to its polar flattening, are easily detected by observing the flight paths of such satellites. Indeed, these satellite studies have shown that the Earth is slightly pear-shaped.

You may have heard of this technique in connection with the various flights of spacecraft around the Moon in recent years. The presence of concentrations of more dense materials at particular places below the Moon's surface—the so-called mascons—was first detected in this way, from the deviations in the flight paths of spacecraft in lunar orbit.

If we take the value

$$Gm_E = 3.986\,026 \times 10^{14}\ \mathrm{m^3\ s^{-2}}$$

with the value of G, already quoted: $G = 6.668 \times 10^{-11}\ \mathrm{N\ m^2\ kg^{-2}}$

we find $m_E = 5.978 \times 10^{24}$ kg

From the known shape and size of the Earth we can calculate its volume and hence its average density.

We find that the *average density of the Earth* = 5 515 kg m^{-3}.

Take note of the last figure—we shall be referring to it later in this Unit.

22.3.3 Movements of the Earth's axis

The Earth does not simply rotate about its own axis and orbit around the Sun. Due to the gravitational attractions of the Sun and the Moon, the Earth's axis of rotation traces out a conical figure in space as shown in Figure 3. This motion is called *precession* of the axis and takes place because the Earth's axis is at an angle (23.5°) to the plane of the ecliptic†, and because the Earth is not exactly spherical. The gravitational attractions of the Sun and the Moon tend to pull the equatorial bulge into the plane of the ecliptic. If the Earth were not rotating, then the equator would simply line up with the ecliptic, in much the same way as a top falls on its side when it stops spinning. It is its rotation that causes the Earth,

precession

instead, to *precess*. For the same reasons the axis of a spinning top moves slowly around in a circle if it is not exactly vertical.

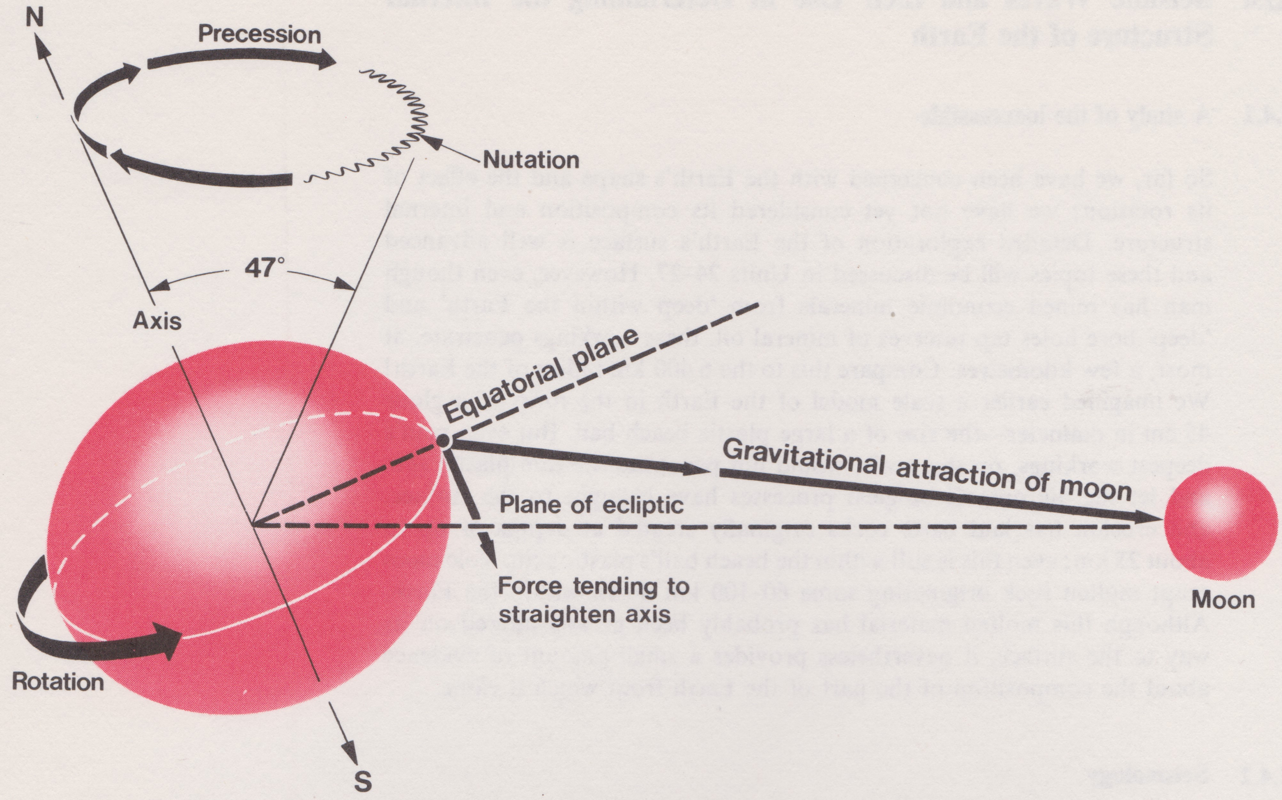


Figure 3 The Earth's precession results from the gravitational attraction applied to the equatorial bulge.

The concept of precession is rather difficult to explain in simple terms, and since this effect is an *external* one we shall not discuss it further in this Unit, which is primarily concerned with the *internal* structure of the Earth.

You may note in passing, however, that the precessional motion of the Earth's axis is rather slow—it takes some 26 000 years to make one precessional revolution. Note also that the gravitational pull of the Sun and the Moon on the equatorial bulge fluctuates according to their position with respect to the Earth. This results in a small wiggle, called *nutation*, being superimposed on the precessional motion.

nutation

There is also another, very much smaller, motion of the Earth's axis which may be due to processes going on *inside* the Earth, and which is accordingly of interest to us in this Unit and in Unit 23.

Imagine you are looking down on the geographic North Pole from space. If you could see the point where the Earth's imaginary axis of rotation emerges from the Earth's surface, that point would be moving anticlockwise in a roughly circular path some 13 m in diameter. The point moves fairly slowly, taking about 14 months to go once around the circle. This motion is called the *Chandler wobble*. It may be brought about by the redistribution of masses within the Earth caused by earthquakes or by variations in the rate of rotation of the Earth. The phenomenon is not yet thoroughly understood; but if you would like to know about some of the latest ideas on the cause of the Chandler wobble, you should read the article by Professor M. A. Chinnery in *Understanding the Earth* (Chapter 6).

Chandler wobble

22.4 Seismic Waves and their Use in Determining the Internal Structure of the Earth

22.4.1 A study of the inaccessible

So far, we have been concerned with the Earth's shape and the effect of its rotation; we have not yet considered its composition and internal structure. Detailed exploration of the Earth's surface is well advanced and these topics will be discussed in Units 24–27. However, even though man has mined economic minerals from 'deep within the Earth' and 'deep' bore holes tap reserves of mineral oil, these workings penetrate, at most, a few kilometres. Compare this to the 6 400 km radius of the Earth! We imagined earlier a scale model of the Earth in the form of a globe 45 cm in diameter—the size of a large plastic beach ball. But even man's deepest workings, on this scale, would not penetrate the thin plastic skin and let the air out! Geological processes have brought to the surface, and erosion has laid bare, rocks originally created at depths of up to about 25 km; even this is still within the beach ball's plastic skin. Volcanoes erupt molten rock originating some 60–100 km down within the Earth. Although this molten material has probably been grossly altered on its way to the surface, it nevertheless provides a small amount of evidence about the composition of the part of the Earth from which it came.

22.4.2 Seismology

Without being able to make direct observations, how can we obtain information concerning the composition and structure of the Earth's interior? The indirect methods used are those of geophysics, a science in which the principles of physics and mathematics are combined with the use of precise measuring instruments to determine the physical properties of the Earth and its interior.

As an example of these types of measurements, and the deductions that can be made from them, let us consider the density of the Earth. We have seen (22.3.2) that the average density works out at a figure of just over 5.5×10^3 kg m^{-3}. In contrast to this, direct measurements show the majority of rocks exposed on the Earth's surface to have a density of between 2.5 and 3.0×10^3 kg m^{-3}.

What do you infer from this?

As these figures are lower than the average, the deep interior of the Earth must have a density higher than 5.5×10^3 kg m^{-3}.

Other measurements—of the variations in the Earth's gravitational field, of the amount of heat that is generated within it, of its magnetic properties —all give useful indirect evidence about the gross structure and composition of the Earth. But the most important and significant evidence comes from seismology—the study of earthquakes and the waves they generate.

22.4.3 Earthquake waves and the study of the Earth's interior

When a pebble is dropped into a pool of water, waves pass not only outwards across the surface from the point where the pebble entered but

also downwards through the water to the bottom of the pool. If the pool were of another liquid, say oil or mercury, then the waves generated by the same pebble would differ from those created in water because the *physical properties* of the three liquids differ. An earthquake produces analogous effects to those of the pebble, for the disturbance of rocks creates shock waves and these not only pass around the surface of the Earth but also through its deep interior. From the way they pass through the Earth and the time they take to do so, data concerning some of the physical properties, particularly *compressibility* and *rigidity*, at various depths can be deduced. An earthquake wave is the only natural phenomenon that passes through the body of the Earth in this way and therefore its study is particularly significant.

A large earthquake will cause damage tens, hundreds and occasionally thousands of kilometres away; this man can see and feel. The same earthquake will be recorded on instruments all over the Earth. Present-day earthquake-recording instruments, *seismometers*, are capable of detecting even the minor tremblings of the Earth which, as they cause no damage, pass unnoticed by the man in the street. It has been estimated that some 150 000 earthquakes are recorded each year; and, if the whole surface of the Earth were covered by a network of seismometers, this figure would probably exceed 1 000 000.

seismometer

Fortunately, only a few of these are as disastrous as, for example, the May 1970 earthquake in Peru. As you will see later in this Unit, it is now well understood *where* earthquakes are likely to occur; and in a later Unit we shall discuss *why* this is so. But so far there is no way of predicting *when* earthquakes will take place. It seems unlikely that earthquake prediction will be achieved for a long time to come, if ever, so the problem of protecting populations from the consequences is essentially a social, economic and technological one. Earthquake prediction and the newer, exciting prospect of earthquake modification are discussed by Professor R. L. Kovach in *Understanding the Earth* (Chapter 23).

In the meantime geophysicists are concentrating on learning as much as they can about earthquakes and the shock waves they generate. Vibrations are produced in solids by a sudden blow or rupture. Correspondingly, earthquakes are produced by volcanic explosions and by the breaking of the brittle rocks of the outer part of the Earth. The cause of most earthquakes is a building up of forces in the rocks until they are strained to breaking point when they suddenly fracture and move. When the break occurs, the pent-up energy, accumulated in the rocks during the long period over which the forces built up, is suddenly released. The pattern of forces changes violently and the energy thereby released radiates from the break in the form of waves.

Both the generation and propagation of earthquake waves therefore depend on how rocks behave when deformed and how they react once the breaking point is reached. So before going further with our study of earthquakes, we must take a closer look at the way materials respond to deforming forces.

22.4.4 The deformation of materials

In our previous discussion on dynamics (Unit 4) it was convenient to assume that the colliding bodies were rigid and that they did not change shape on impact. This is, of course, a highly idealized situation and does not correspond to reality. Materials differ greatly in their ability to recover their original shape once the force is removed. A hard rock is difficult to deform whereas a rubber ball or a lump of putty is not. When the force is removed, the rubber ball regains its former shape, but the lump of

putty does not. A body that returns completely to its former size and shape is said to be *perfectly elastic*; one that retains completely its altered size and shape is said to be *perfectly plastic*. Once again these are idealizations; the behaviour of real materials is to be found somewhere between these limiting cases.

perfect elasticity

perfect plasticity

For bodies that are close to being perfectly elastic it is found that, in addition to being able to regain their original dimensions on removal of the force, they have the property of deforming to an extent that is proportional to the applied force; when a material exhibits this behaviour it is said to obey *Hooke's Law*.

Hooke's Law

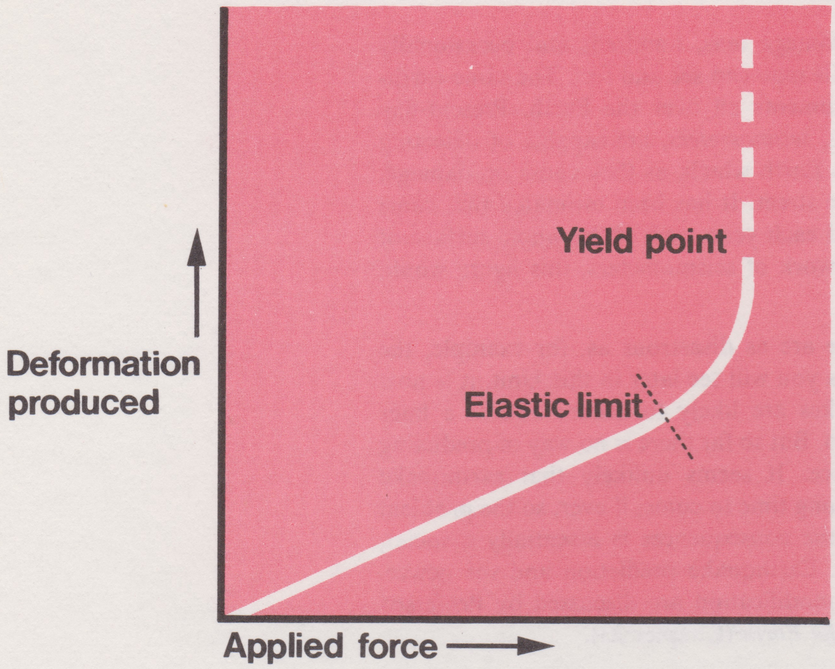


Figure 4 The behaviour of deformed materials.

But, as the force is increased, even with the most elastic substances, there comes a stage where Hooke's Law no longer holds. This point is called the *elastic limit* of the material. It can be found by plotting the extent of the particular type of deformation being studied against the applied force causing it as in Figure 4. Beyond the elastic limit the graph departs from a straight line. Alternatively, the elastic limit may be defined as the magnitude of the maximum force that can be applied to a specimen and the specimen still be able to recover its original dimensions. For all practical purposes the value for the limit comes out the same regardless of which of these two ways we choose to determine it.

elastic limit

Having passed the elastic limit, further increase in the applied force brings us eventually to a point where a fracture occurs; this we call the *yield point*. A change in the dimensions of a body is called a strain.* There are various kinds of strain depending upon the exact nature of the applied forces. One type that is particularly easy to demonstrate is that produced in a wire or length of rubber clamped at one end and supporting a weight at the other (Fig. 5). If the original length of the specimen is l and the extension produced is e, the strain for this type of deformation is defined as:

yield point

$$\text{strain} = \frac{e}{l}$$

* *The terms* strain *and* stress *have a technical meaning here, which you should not confuse with the various other meanings they have in everyday language.*

For a given weight—that is, force—the extension produced will depend upon the thickness of the specimen. So it is useful to define a quantity called stress, which is the force applied per unit area. In the case we are considering, the stress is the stretching force exerted by the weight divided by the cross-sectional area of the specimen of wire. It is called a longitudinal stress as it acts in the direction of the length of the specimen. Hooke's Law can be expressed in the general form:

$$\frac{\text{stress}}{\text{strain}} = \text{constant}$$

and in the case of a longitudinal stress this becomes:

$$\frac{\text{longitudinal applied force/cross sectional area}}{\text{change in length/original length}} = Y \quad \ldots\ldots(2)$$

Y is a constant called *Young's modulus* and it is a characteristic of the material of the specimen. (The word 'modulus' simply means 'constant factor'.)

Young's modulus

When a length of rubber is stretched what else do you notice about its dimensions besides its increase in length?

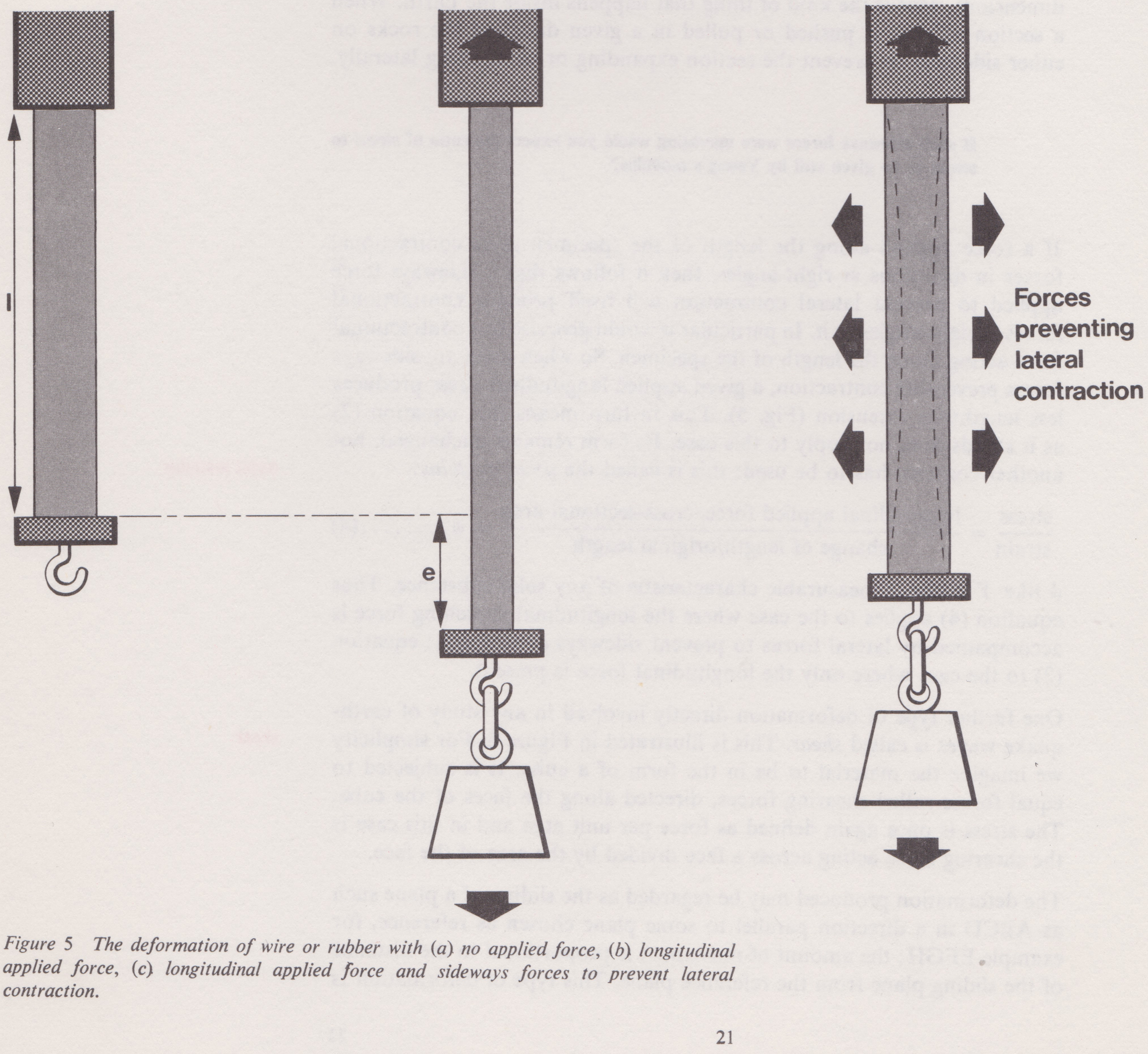


Figure 5 The deformation of wire or rubber with (a) no applied force, (b) longitudinal applied force, (c) longitudinal applied force and sideways forces to prevent lateral contraction.

When a specimen is stretched, not only does its length increase but its width is decreased. It is found that this lateral change in dimensions is proportional to the longitudinal strain:

$$\frac{\text{lateral strain}}{\text{longitudinal strain}} = \frac{\text{change in width/original width}}{\text{change in length/original length}} = \sigma \text{ ....(3)}$$

σ is a constant for the material of the specimen and is called *Poisson's ratio*. **Poisson's ratio**

The equations describing elastic behaviour strictly apply only to cases where the forces and strains are small, so it essentially makes no difference whether we use the cross-sectional area before or after stretching in equation (2). Incidentally you should note that equations (2) and (3) apply equally to the case where the longitudinal force is compressive—then we get longitudinal contraction and lateral expansion.

You might like to take this opportunity of doing home experiment 22.1. In it you can test Hooke's Law and measure Young's modulus for a specimen of rubber.

Since stress may be applied to a body in various ways, giving rise to various types of strain, a material possesses a number of other elastic moduli which we shall now consider. For example, we can imagine a system of applied forces that gives a longitudinal strain only. This we could achieve by applying a sideways force to maintain the original lateral dimensions. This is the kind of thing that happens inside the Earth. When a section of rock is pushed or pulled in a given direction the rocks on either side tend to prevent the section expanding or contracting laterally.

If such sideways forces were operating would you expect the ratio of stress to strain to be given still by Young's modulus?

If a force applied along the length of the specimen gives contractional forces in directions at right-angles, then it follows that a sideways force applied to prevent lateral contraction will itself produce contractional forces at right-angles to it. In particular it would give rise to a contractional force acting along the length of the specimen. So when there are sideways forces preventing contraction, a given applied longitudinal stress produces less lengthwise extension (Fig. 5). This in turn means that equation (2) as it stands does not apply to this case. Its form remains unchanged, but another constant has to be used; this is called the *axial modulus*: **axial modulus**

$$\frac{\text{stress}}{\text{strain}} = \frac{\text{longitudinal applied force/cross-sectional area}}{\text{change of length/original length}} = \psi \text{ ....(4)}$$

ψ like Y is also a measurable characteristic of any solid substance. Thus equation (4) applies to the case where the longitudinal stretching force is accompanied by lateral forces to prevent sideways contraction; equation (2) to the case where only the longitudinal force is present.

One further type of deformation directly involved in any study of earthquake waves is called *shear*. This is illustrated in Figure 6. For simplicity we imagine the material to be in the form of a cube. It is subjected to equal forces called shearing forces, directed along the faces of the cube. The stress is once again defined as force per unit area and in this case is the shearing force acting across a face divided by the area of the face. **shear**

The deformation produced may be regarded as the sliding of a plane such as ABCD in a direction parallel to some plane chosen as reference, for example EFGH; the amount of movement is proportional to the distance of the sliding plane from the reference plane. This type of deformation is

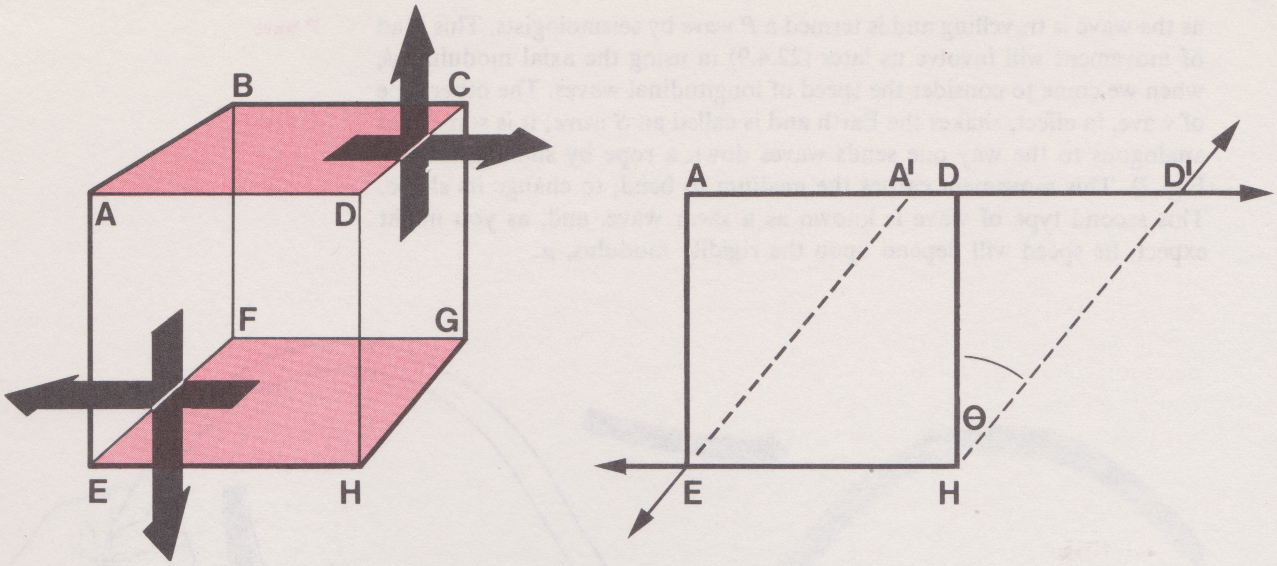


Figure 6 Shear deformation.

called *shear strain* and is illustrated in Figure 6 (b) where the corners A and D of the cube have moved to new positions A′ and D′. **shear strain**

Does shear strain involve a change in volume? Look again at Figure 6 (b).

From simple geometrical considerations (see *MAFS*, section 2) the area ADHE is equal to the area A′D′HE. This, together with the fact that distances in the third dimension (for example, DC in Fig. 6) are also unchanged, means that the volume of the cube is unaltered; so shear strain produces a change in *shape only*—not in size. This type of strain is expressed quantitatively as the displacement suffered by a plane divided by the distance separating this plane from the reference plane. For small displacements (and remember the theory of elasticity applies only to small displacements) this becomes approximately equal to the angle θ in Figure 6 (b).

$$\text{Thus, shear strain} = \frac{\text{relative displacement of two planes}}{\text{separation between planes}}$$

$$= \frac{DD'}{DH} = \tan\theta \approx \theta \text{ (in radians) (See } MAFS\text{, section 4)}$$

For small displacements Hooke's Law is once again obeyed; this time

$$\frac{\text{stress}}{\text{strain}} = \frac{\text{shearing force/unit area}}{\text{angular deformation}} = \mu \quad \dots\dots\dots\dots\dots\dots(5)$$

μ is a measurable constant for the material and is called the *rigidity modulus*. **rigidity modulus**

What kind of value would you expect μ to have for water and for air?

For future reference you should note that μ is always zero for a liquid or gas because these media are completely unable to resist a shearing stress.

As we shall see later, two distinct types of wave can pass through the Earth. One is a *longitudinal wave*, the other a *shear wave*. The first type produces successive compressions and rarefactions (refer back if necessary to our discussion on the nature of sound in Unit 2) in the same direction

as the wave is travelling and is termed a *P wave* by seismologists. This kind of movement will involve us later (22.4.9) in using the axial modulus, ψ, when we come to consider the speed of longitudinal waves. The other type of wave, in effect, shakes the Earth and is called an *S wave*; it is somewhat analogous to the way one sends waves down a rope by shaking it* (see Fig. 7). This movement causes the medium to bend, to change its shape. This second type of wave is known as a shear wave, and, as you might expect, its speed will depend upon the rigidity modulus, μ.

P wave

S wave

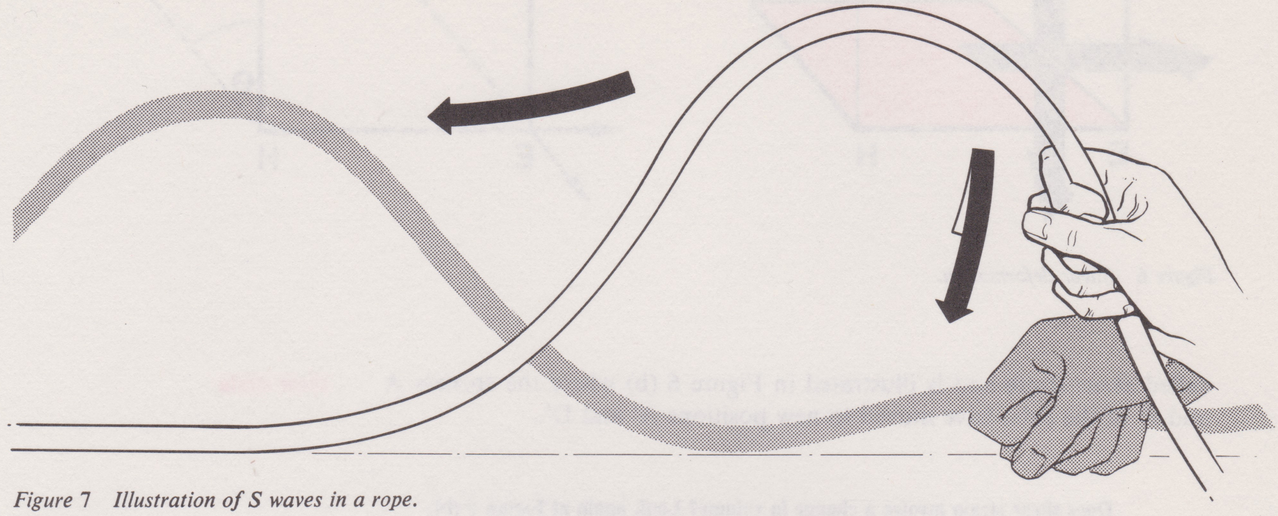


Figure 7 Illustration of S waves in a rope.

22.4.5 Earthquake mechanisms

Nearly all earthquakes are produced by shearing forces, so that when the yield point of the rock is reached, the Earth fractures and all adjacent points move laterally apart as shown in Figure 8. When this happens, the two kinds of waves (P and S) are sent out from the place where the fracture has occurred.

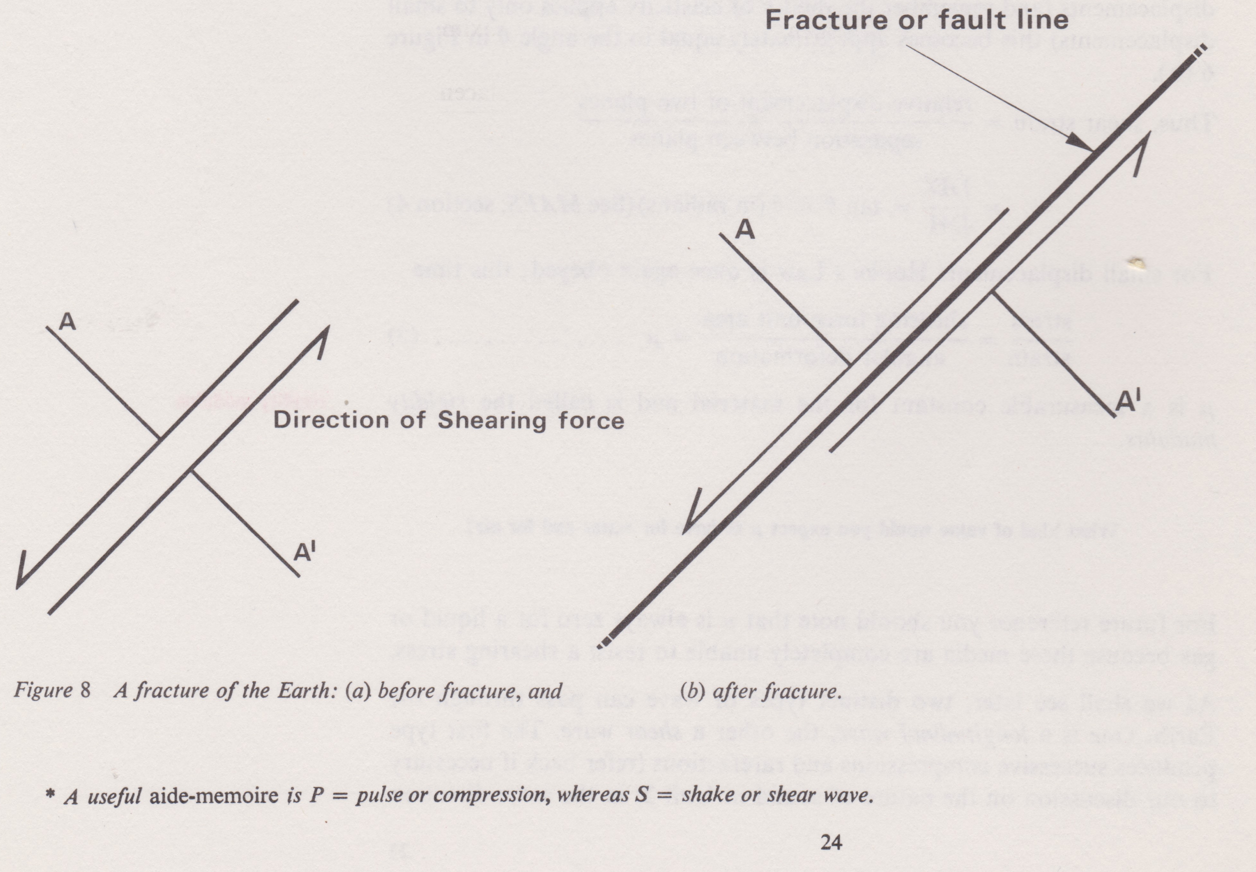


Figure 8 A fracture of the Earth: (a) before fracture, and (b) after fracture.

* *A useful* aide-memoire *is P = pulse or compression, whereas S = shake or shear wave.*

At the moment of fracture, the rock to the right of the fault in Figure 8 will push, or compress, the adjacent rock in the direction of motion (top right), and pull away from, or decompress, the rock in the opposite direction (bottom right). But on the other side of the fault, the rock moves in the opposite sense; and so it again compresses the adjacent rock in the direction of motion (bottom left) and decompresses the adjacent rock in the opposite direction (top left).

The result of these opposing motions is that P waves are not transmitted along the fault as you might expect them to be. At some distance from the fault, in the direction of the fault itself, the opposite movements cancel out and no longitudinal or P wave is transmitted in this direction. Nor, by definition, are longitudinal waves transmitted at right angles to the fault because there is no rock movement in that direction. They are, in fact, transmitted in all other directions, so that the direction of maximum strength of the P shock wave lies at 45° to the fault (Fig. 9).

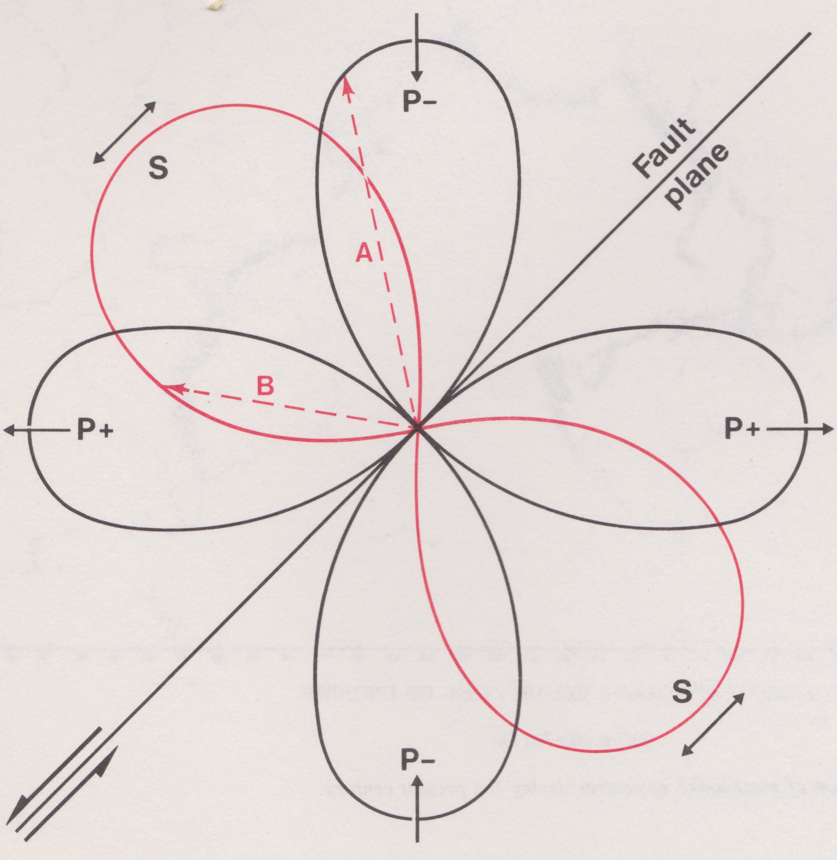


Figure 9 The pattern of P and S waves produced at an Earth fracture. The intensity of the waves emitted in any given direction is proportional to the length of lines such as A (for P waves) and B (for S waves). Thus line A has a maximum length—and thus P waves have maximum intensity—at 45° to the fault. Similarly, line B has a maximum length—and thus S waves have maximum intensity—at 90° to the fault. The intensity of P waves along the fault and at 90° to the fault is zero (line A has zero length); and the intensity of S waves along the fault is zero (line B has zero length).

The shear (S) waves, on the other hand, will be strongest in the direction at right angles to the fault motion and zero along the fault. The directional characteristics of P and S waves propagated from an earthquake are shown in Figure 9.

22.4.6 Earthquake zones

Most earthquakes originate from fractures occurring very close to the surface of the Earth, usually at depths less than 30 km. Some, however, have been recorded at depths down to 700 km.

The majority of earthquakes originate in well-defined *earthquake or seismic zones*. In Figure 10 you can see that there are two main continental seismic zones, one bordering the Pacific and a broader zone corresponding roughly to the Alpine-Himalayan mountain chain. These are both mountainous regions where the outer part of the Earth is being actively deformed. You can also see that seismic belts occur in the oceans; these again are associated with mountainous areas, submarine ranges whose peaks are hundreds of metres below sea-level.

seismic zone

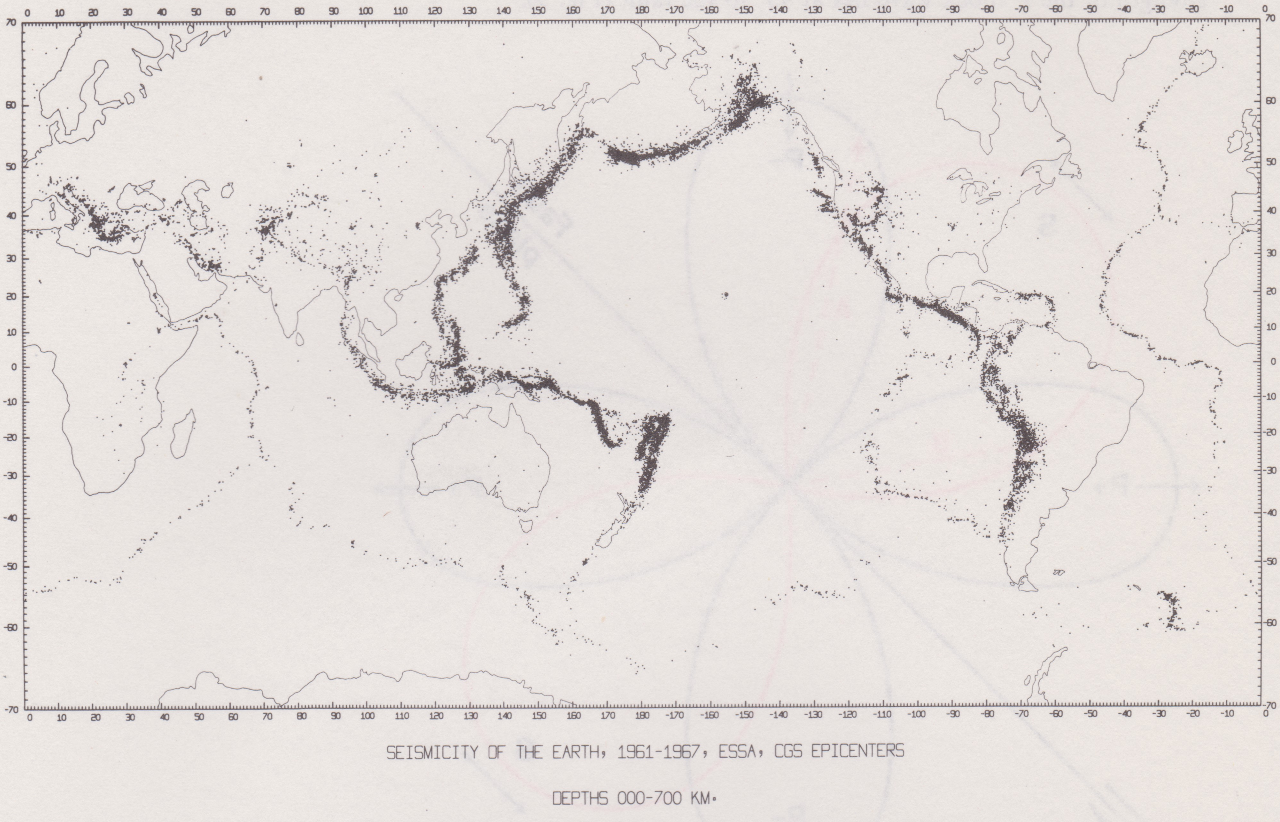


Figure 10 *Map showing the distribution of earthquake epicentres during the present century.*

Most earthquakes are caused by the fracturing of the Earth along an extensive fault zone. Therefore the waves produced emanate from a planar source rather than a point source. However, it is easier to make deductions if the waves are considered as coming from a point source. Provided that the waves are being recorded at a distance which is large compared with the size of the fault, it is sufficiently accurate for our purposes to treat the earthquake as a point source. This 'point source' of earthquake waves is known as the *focus* of the earthquake. The focus can be deep within the Earth and it is useful to have some surface reference point—a line drawn vertically from the focus cuts the Earth's surface at a point known as the *epicentre*.

earthquake focus

epicentre

22.4.7 Earthquake magnitude and intensity

Earthquakes vary from mild quaverings to violent oscillations of the Earth's surface. The magnitude of an earthquake is directly related to

the amount of energy released when over-strained rocks fracture. The greater the stress at which this occurs, the greater the energy released and the greater the magnitude of the earthquake. So the *magnitude* depends essentially on the inherent breaking strength of the rocks in which the fracture occurs. **earthquake magnitude**

The *intensity* of the vibrations due to an earthquake is expressed in terms of an arbitrary, 12-point scale which is essentially descriptive of the damage done at some point on the Earth's surface. The scale goes from I: 'Instrumental' (only detectable by instruments), to XII: 'Catastrophic'. The intensity decreases with increasing distance from the focus. A line drawn through all places which suffer the same intensity is known as an *isoseismal line*; the relationship between focus, epicentre, intensity and isoseismal line is shown on Figure 11. **earthquake intensity** **isoseismal line**

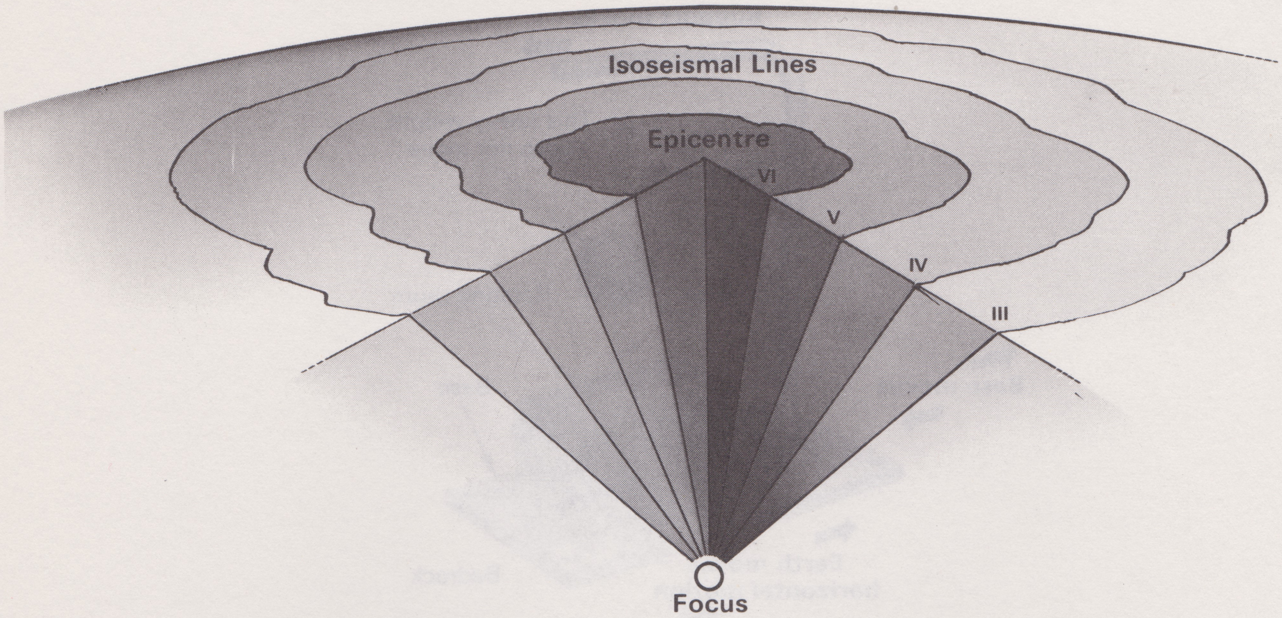


Figure 11 Diagram showing the focus of an earthquake, the epicentre and isoseismal lines. The Roman numerals indicate intensities.

When an earthquake occurs it is rarely a simple break sending out a brief pulse of waves. Sometimes, when the yield point of the rocks is reached, small incipient slips or fractures will occur first. Then the main movement takes place and finally there are often reverse readjustment movements because the main movement has over-reacted to the applied stress. Chronologically these are termed the *foreshocks*, the *principal shock* and the *aftershocks*. One difficulty in interpreting earthquake wave signals is therefore that the parcel of waves from an earthquake is a complicated one. Another difficulty is that some of these waves travel through the Earth and others through the rocks near the surface. You already know that there are two types of wave passing through the Earth, longitudinal and shear. Later, in section 22.4.9, we shall see that the longitudinal waves travel about twice as fast as the shear waves. This being so, the time interval between the arrival of the two types of waves will increase with distance travelled. **foreshock** **principal shock** **aftershock**

22.4.8 Recording of earthquake waves

The task of the seismometer at the recording station is to detect vibrations of the Earth which may be as small as 10^{-6} m. A recording station will usually have 3 seismometers—one to pick up vibrations in the vertical

direction and two others to record in an east-west and north-south direction, all three being known as a seismic set. Modern practice is to have as many as 100 seismometers laid out in an L- or a T-shaped arrangement; such a group of instruments is known as a seismic array. The principle on which the seismometer works is that part of it remains stationary while the waves pass whereas the rest of the apparatus, in direct contact with the ground, oscillates. The magnitude and frequency of the oscillations are then recorded on a time-correlating device; this record is a *seismogram*. Modern seismometers are electronic devices which detect, amplify, filter and record the motions of the Earth, but the principle is exactly the same as that illustrated in Figure 12.

seismogram

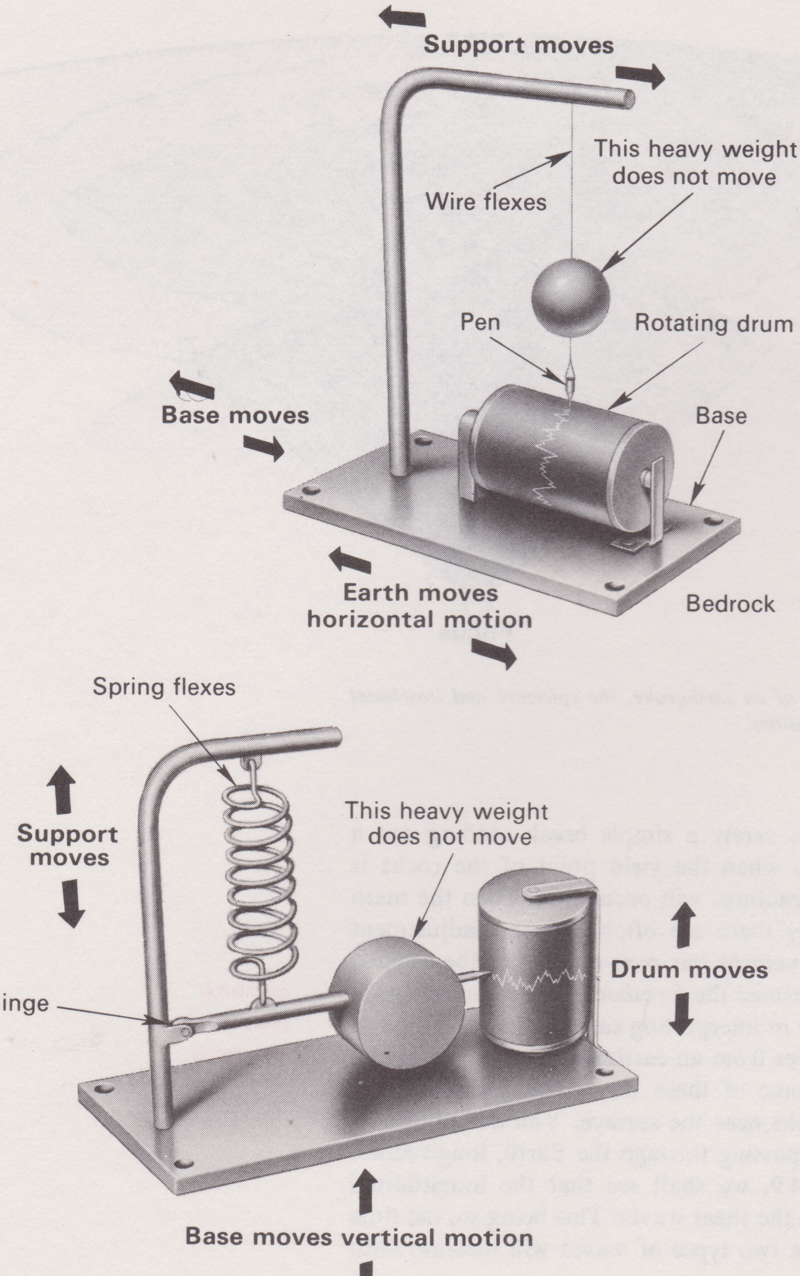


Figure 12 The principle of the seismometer. Horizontal motions are detected by the mechanical arrangement shown at the top, and vertical motions by the arrangement shown at the bottom.

22.4.9 Wave propagation

We have considered in general terms processes by which earthquake waves are created and recorded. However, before we can arrive at our main objective—to investigate the structure and composition of the Earth's interior—we must explain how earthquake waves are propagated through the Earth.

The velocity of a wave depends upon the properties of the medium through which it passes. If we measure the wave velocity, therefore, we can deduce something about the nature of the medium. The relationships between the velocities of both longitudinal and shear waves and the properties of the transmitting medium are, in fact, extremely simple. We shall give them here without proof; but if you are interested in seeing how they are derived you should turn to Appendix 2 (Black).

For a longitudinal wave:

$$\text{velocity} = \left(\frac{\text{axial modulus of medium}}{\text{density of medium}}\right)^{\frac{1}{2}}$$

$$\text{or, } u = \left(\frac{\psi}{\rho}\right)^{\frac{1}{2}} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots(6)$$

And for a shear wave:

$$\text{velocity} = \left(\frac{\text{rigidity modulus of medium}}{\text{density of medium}}\right)^{\frac{1}{2}}$$

$$\text{or, } \omega = \left(\frac{\mu}{\rho}\right)^{\frac{1}{2}} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots(7)$$

There is an important consequence of equation (7). Because, for a liquid (or a gas), $\mu = 0$, it immediately follows that $w = 0$. In other words, *liquids (and gases) are incapable of transmitting shear waves.* This result is essential to our study of the structure of the Earth, as you will soon learn.

For waves passing through the Earth, the velocity of longitudinal waves varies between about 8 and 14 km s^{-1}; and shear waves have approximately half this velocity—that is:

$$\mu \approx 2w.$$

What, then, is the ratio of ψ to μ for the materials of the Earth's interior?

$u = \left(\frac{\psi}{\rho}\right)^{\frac{1}{2}}$ and $w = \left(\frac{\mu}{\rho}\right)^{\frac{1}{2}}$

so $u^2 = \frac{\psi}{\rho}$ and $w^2 = \frac{\mu}{\rho}$

therefore $\frac{u^2}{w^2} = \frac{\psi/\rho}{\mu/\rho} = \frac{\psi}{\mu}$

So if, $u \approx 2w$

$u^2 \approx 4w^2$

and $\frac{\psi}{\mu} = \frac{u^2}{w^2} \approx 4$

So, generally speaking, the longitudinal waves generated by an earthquake can be expected to arrive at a recording station before the shear waves. For this reason seismologists call the former primary waves or *P waves* and the latter secondary or *S waves.*

22.4.10 General principles governing wave propagation

As you will see later, the seismologist does not measure the velocities of P and S waves. All he can measure is the *time* it takes for a wave to go from the earthquake to the detecting station. To deduce the *velocity* (and hence something about the density and the elastic moduli) from the *time*, he must know the distance the wave has travelled through the Earth's interior. In the Earth the elastic properties of the medium alter from one region to another. So the wave velocities will also vary from one region to another; and at the boundaries between these regions the waves will also be reflected or refracted, in a manner which we shall shortly describe. As a

result of reflection and refraction the waves may not be propagated from earthquake to detecting station along simple, straight-line paths. The wave paths, or rays, may be curved, or they may zig-zag through some parts of the Earth owing to multiple reflections. If we want to deduce elastic properties from velocity and velocity from time of travel, we must be sure about the distance travelled. And if this may depend upon reflection and refraction, we have to be in a position to take these effects into account.

So this is the moment to set aside our study of earthquake waves in order to discuss the laws of reflection and refraction of waves in general. These depend upon two quite general principles which apply to any sort of wave propagation—earthquake waves, sound waves, light waves—any waves at all.

The first is called the *superposition principle*. This applies when two or more waves cross the same region of space. Consider, for example, ripples spreading out over the surface of water from two similar sources. In the region where the ripples overlap there are places where the disturbance is practically zero and others where the disturbance is about twice as great as it would be if there were only one source.

superposition principle

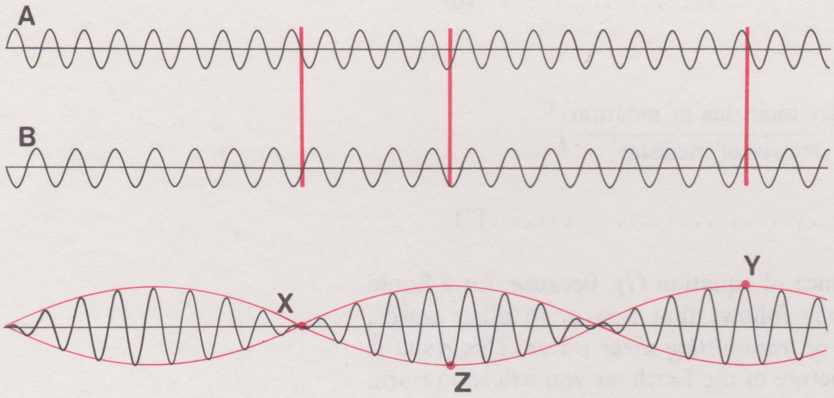


Figure 13 *Illustration of the superposition principle.*

The *superposition principle* states that the resultant amplitude (see Unit 2) at a point is the sum of the amplitudes of the various waves at that point. For example, waves A and B in Figure 13 add up to the resultant shown. At point X the crest of one wave train coincides with the trough in the second so if the amplitudes of the individual waves are equal they cancel and produce a resultant displacement of zero. At the points Y and Z the wave trains are in step and produce an amplitude that is twice that of either wave A or B considered separately. This superposition of wave trains is called *interference*. At point X where a cancellation occurs there is *destructive interference*. At Y and Z where there is an enhancement there is *constructive interference*.*

interference

The fact that amplitudes can simply be added means that once the waves have passed through the interference zone they continue unaffected—it is as though they had passed through undisturbed space. An example of this is shown in Figure 14 where you see a photograph of ripples spreading out over the surface of water—having crossed other ripples they carry on unaffected.

* *Note that in order to produce interference, at least some part of the amplitudes of the waves must lie in the same direction—if the amplitudes are at right angles they cannot affect each other.*

Can you think of an example of this involving light waves?

Actually you need look no further than this page for evidence that the superposition principle applies to light waves. The fact that you are able to read this print means that the light waves travelling from the print to your eyes are unaffected by whatever other stray light is present in the room and is passing through the intervening space.

The second general principle is *Huygens' principle*, so named after a Dutch scientist who played a prominent part in the formulation of the wave theory of light. It is illustrated in Figure 15 which shows water waves advancing towards a barrier which has a small hole in it. The barrier

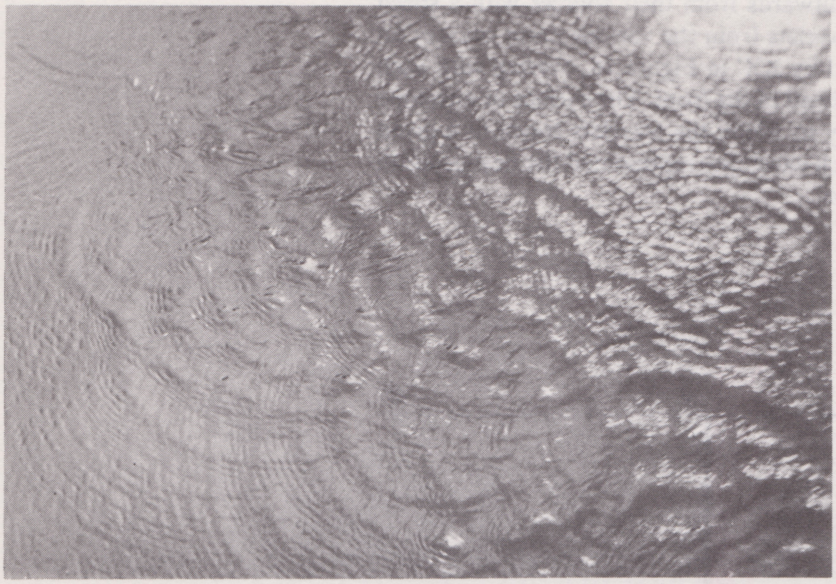


Figure 14 Photograph of ripples produced by raindrops, showing how the waves cross through each other.

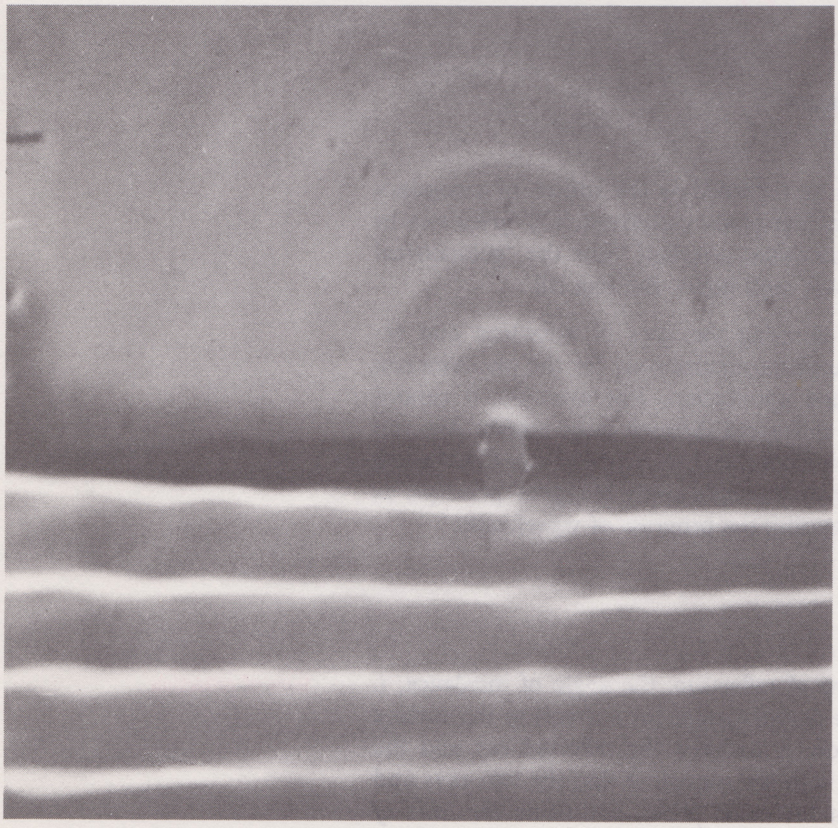


Figure 15 Photograph of water waves passing through a small hole.

prevents all but a small part of the wave front from continuing towards the bottom of the picture. This arrangement allows us to examine the behaviour of an isolated portion of the front. You see the wave spreading out beyond the barrier. This type of behaviour, as you learnt in Unit 2, is called diffraction. It is as though the portion of the wave front passing through the hole were acting as a source of secondary wavelets spreading out in all directions. This is the essence of Huygens' principle; *it states that each point on a wave front acts as though it were a source of secondary wavelets.*

Huygens' principle

The wavelets progress radially outwards from their sources with a speed characteristic of the medium. Wavelets from the various point sources on a wave front like that represented by XX′ in Figure 16, overlap with each other in the region immediately ahead of the front. The surface or envelope YY′ marks the furthest extremity to which the wavelets travel in a given time. Huygens postulated that such envelopes represented the new position of the wavefront at this later time.

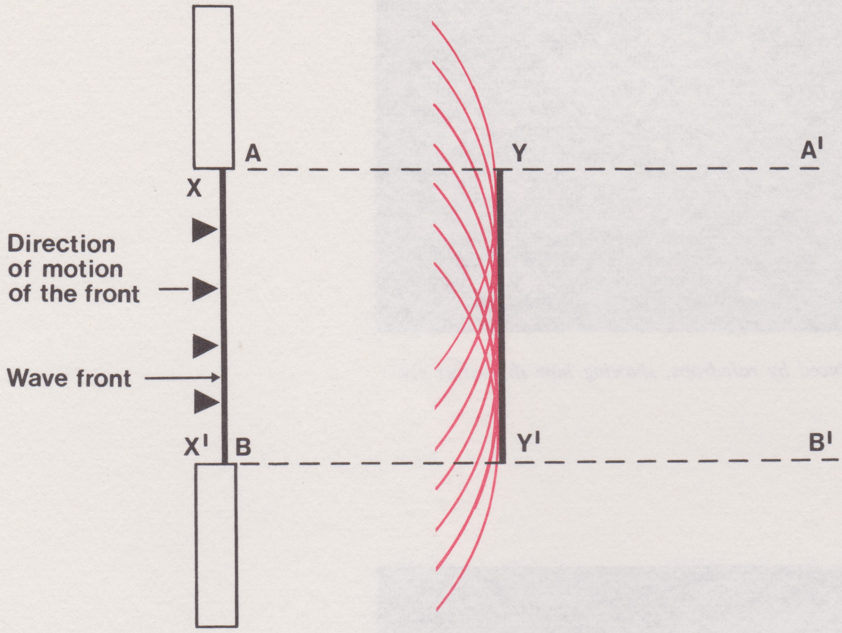


Figure 16 *Illustration of Huygen's principle.*

This new wavefront can in turn be regarded as the source of secondary wavelets and so on. You see that the plane wavefront XX′ gives rise to another wavefront YY′ that somewhat overlaps into the 'geometrical shadow' that is, above the line AA′ and below the line BB′.

Huygens' principle is useful in deriving the laws of reflection and refraction. It can also be used to account qualitatively for the phenomenon of diffraction, but it is of limited application. For example, it cannot specify the intensity of the light diffracted through large angles. The full explanation of the principle lies much beyond the scope of our Course and we shall pursue it no further. We shall simply apply Huygens' principle as a useful geometric method of mapping the progress of wavefronts.

22.4.11 Reflection and refraction

It is well known that, in the case of reflection, the angle of incidence, i, between the incident ray and the perpendicular (or *normal*) to the surface is equal to the angle of reflection, r, between the reflected ray and the normal (Fig. 17). This, together with the statement that the two rays and the normal to the surface lie in the same plane, constitute *the laws of reflection*. With the help of Huygens' principle we can now derive these

laws of reflection

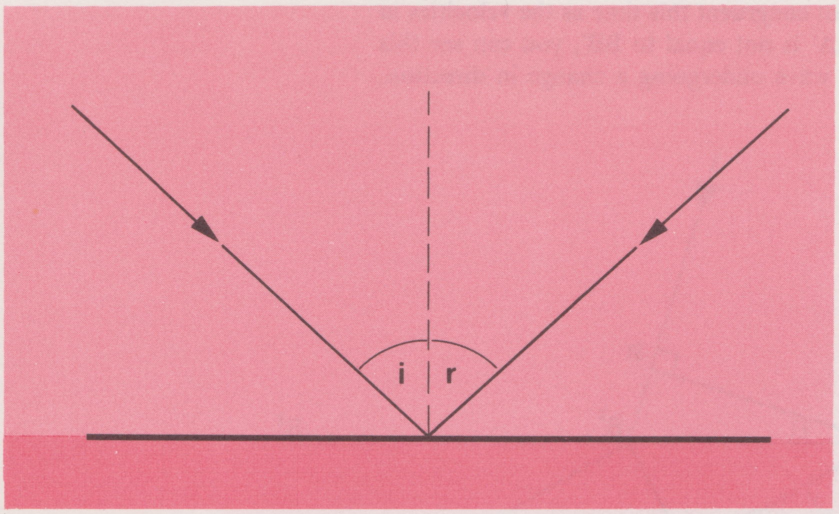


Figure 17 *The angles of incidence and reflection.*

laws. In Figure 18 a wavefront AB strikes a plane SS′ (the lines drawn perpendicular to the wavefront are rays of light; they indicate the direction in which the front is moving). The moment point A on the front makes contact with the plane it emits secondary wavelets in all directions. After

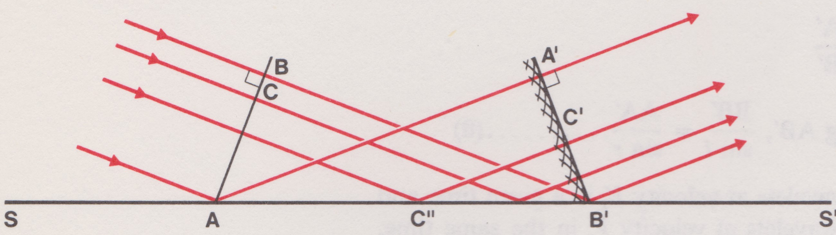


Figure 18 *Reflection of waves from a plane surface.*

a given time interval, B strikes SS′ at a point B′ and secondary wavelets begin to spread out from B′. By this time the wavelets from A have reached as far as A′. We also show an intermediate point C giving rise to wavelets emanating from C″ and reaching as far as C′ by the end of the interval. The wavelets are found to be concentrated along the line A′C′B′; this envelope is the new position of the wavefront. The direction of propagation of the wave after reflection is perpendicular to this wavefront. The time taken for the wave to go from B to B′ is the same as that required to go from A to A′.

Therefore: $\quad BB' = AA'$

Triangles ABB′, B′A′A are congruent (see *MAFS*, section 2).

Therefore: $\quad BB'A = A'AB'$

In other words the angle of incidence equals the angle of reflection (Fig. 17).

Suppose now the wave can pass across a boundary from one medium into another. Let the wave velocities in the two media be V_i and V_r. In Figure 19 the wave front AB strikes the boundary SS′. The point A on the front arrives at the boundary and emits secondary wavelets. These reach point A′ by the time B has reached B′. Line A′B′ marks the envelope where the wavelets are concentrated and so represents the new wavefront.

Triangles AA′B′ and B′BA are not congruent this time as the velocities in the two media differ. Because AA′ is not equal to BB′, you can see this change of velocity results in the wave undergoing a change in direction.

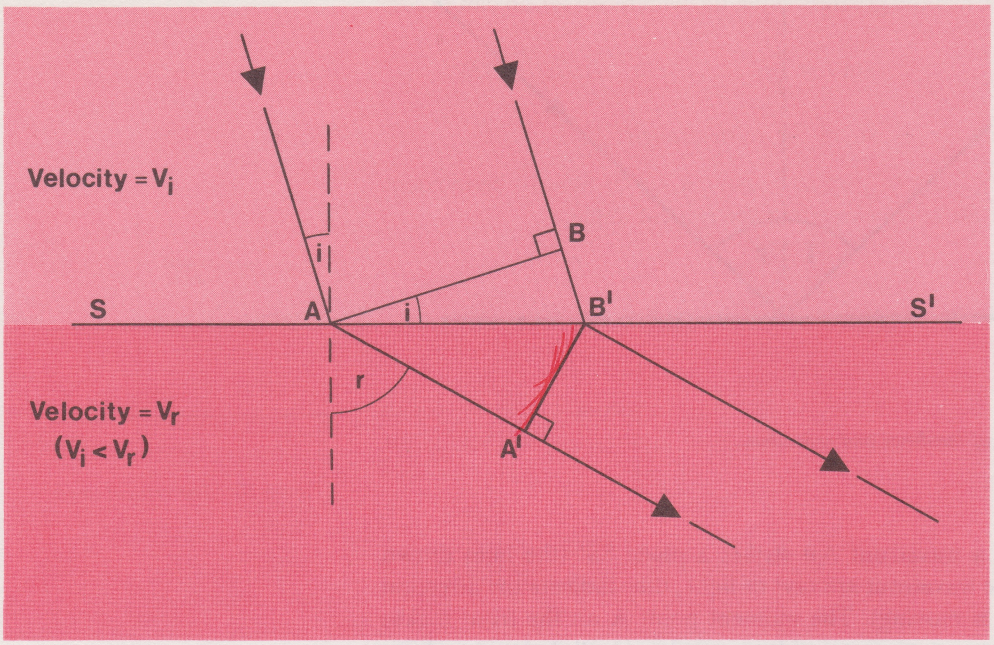


Figure 19 *Refraction at a plane surface.*

Now $\sin i = \dfrac{BB'}{AB'}$ and $\sin r = \dfrac{AA'}{AB'}$

Therefore: eliminating AB′, $\dfrac{BB'}{\sin i} = \dfrac{AA'}{\sin r}$ ....(8)

BB′ is the distance travelled by wavelets at velocity V_i in a given time and AA′ is the distance travelled by wavelets at velocity V_r in the same time.

Therefore: $\dfrac{BB'}{AA'} = \dfrac{V_i}{V_r}$ ....(9)

From equations (8) and (9):

$$\frac{\sin i}{\sin r} = \frac{V_i}{V_r} \quad \text{....................(10)}$$

For two given media V_i/V_r is a constant. Thus the *angle of refraction*, *r*, (that is, the angle the refracted ray makes to the normal to the surface) is related to the angle of incidence by the expression:

$$\frac{\sin i}{\sin r} = \text{constant} \text{....................(11)}$$

Equation (11) is known as *Snell's Law*, after its discoverer.

Snell's Law

Note that when a wave passes from a medium of high velocity to one of lower velocity, it is refracted towards the normal to the surface; when passing from a slower medium to a faster one, it is refracted away from the normal. If the incident ray is a light ray travelling in a vacuum (or for all practical purposes in air), the constant in equation (11) is a characteristic of the medium into which the ray passes, and is called the *refractive index*. It is denoted by the letter *n*.

refractive index

Therefore: $\dfrac{\sin i}{\sin r} = n$ ....(12)

Equation (11) together with the statement that the incident ray, refracted

ray, and the normal to the surface lie in the same plane, constitute *the laws of refraction*.

You can see from equation (12) that when $r = 90°$ (that is, $\sin r = 1$) the angle of incidence, i, has the greatest value for which refraction can occur.

Then: $$\sin i = \frac{V_i}{V_r} \quad \text{(13)}$$

This particular value of i is called the *critical angle*.

critical angle

There is of course nothing to stop us causing a ray of light to be incident on the boundary at an angle exceeding the critical angle.

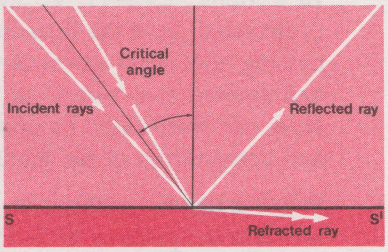


Figure 20 The critical angle.

What do you expect would happen to such a ray?

When this happens the light cannot refract into the second medium—it can only reflect back into the original medium, as in Figure 20. Such behaviour is called *total internal reflection*.

total internal reflection

Would you expect total internal reflection to be possible on both sides of a boundary?

No, it only affects waves trying to pass from a medium of lower velocity to one of higher velocity. If the wave approaches from the opposite side of the boundary there is no angle of incidence for which refraction cannot take place, so one does not get total internal reflection. Total internal reflection plays an important part in seismology. An earthquake wave originating in a layer where the velocity is low can find itself trapped within that layer; it is repeatedly reflected between, say, the Earth's surface above, and the boundary with the next layer below (Fig. 21).

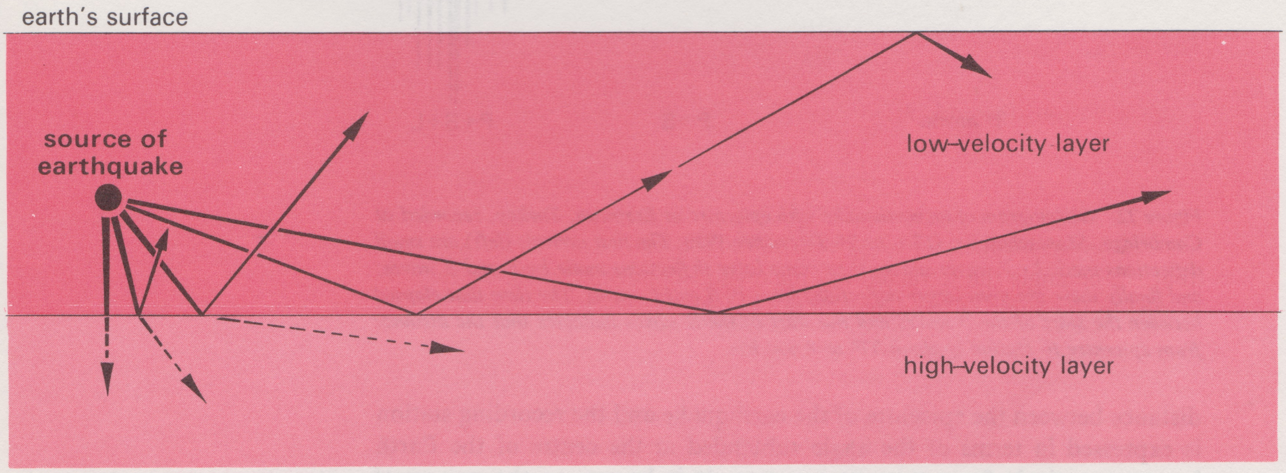


Figure 21 The trapping of waves in a low-velocity layer.

22.5 The Internal Structure and Composition of the Earth

22.5.1 Seismographic data

In sections 22.4.9–11, we showed how longitudinal and shear waves are propagated, gave the formulae for the velocity of propagation (noting, in passing, that shear waves cannot be transmitted through liquids or gases), and showed how waves can be reflected or refracted in different media. How can these ideas be applied to find out about the internal structure and composition of the Earth?

Suppose that an earthquake occurs at some point near the Earth's surface, and that the position of its focus and epicentre is known.* At a recording station somewhere on the Earth's surface arrival times of the various waves from the earthquake can be measured (Fig. 22). When using these data, the

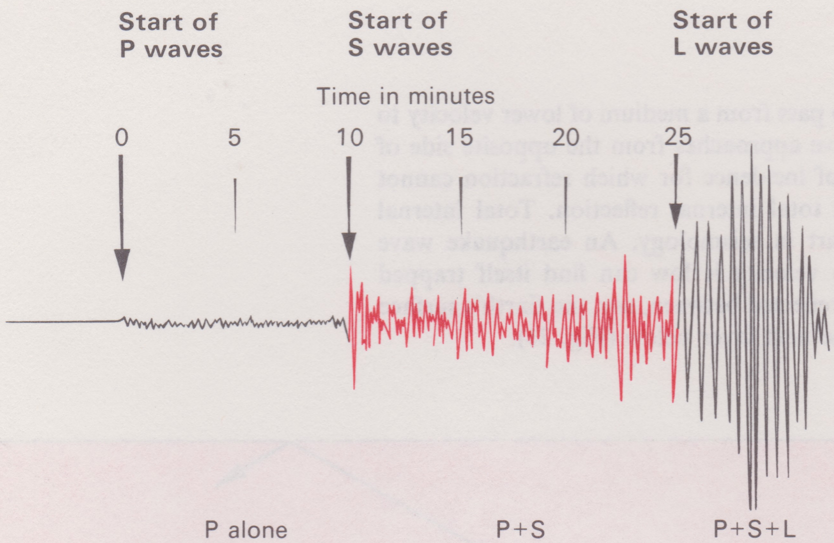


Figure 22 Seismogram of an earthquake with epicentre at Erzincan, Turkey, recorded at Cambridge, Massachusetts, USA, on 26 December 1939. The first arrivals (left) are small deflections made by P waves alone; then come large deflections made by P and S waves; and finally there are even larger deflections from P, S and L waves. The time that elapsed between the start of the P waves and the start of the S waves indicates that the distance from epicentre to station is about 77° or 8 550 km.

distance between the epicentre of the earthquake and the recording station is expressed in terms of the angle subtended at the centre of the Earth between verticals from these two points. It is known as the *epicentral angle* and is the way of expressing this distance. The relationship between distance as measured on the surface of a sphere and the angle that this distance represents at the centre is graphically illustrated in Figure 23.

epicentral angle

* *The trembling of the Earth caused by an earthquake will be felt by people living nearby. By constructing an isoseismal map (Fig. 11) the position of the epicentre can be determined. Furthermore, the spacing of the isoseismal lines will indicate the depth of the focus. The time at which it occurs will also be known. Obviously, this only applies to inhabited areas. The position of the focus and the epicentre and also the time at which the earthquake happened, can, however, be determined by comparing the seismograms from several recording stations. The method by which this is done falls outside the scope of the Course.*

Figure 22 shows a typical seismogram of an earthquake recorded at a station about 8 550 km from the epicentre. It shows first small deflections lasting some 10 minutes, then larger ones lasting about 15 minutes. The P waves travel at about twice the velocity of the S waves and therefore arrive first. Then come the S waves. The very much larger deflections which appear last of all are due to the waves which have travelled around the surface of the Earth; the letter L is used to designate them. It is the P and S or *body* waves which interest us here, as they have travelled through the interior of the Earth. It is possible to identify the P and S waves on a seismogram because the individual seismometers are orientated to distinguish between the vertical and horizontal oscillations of the Earth.

body waves

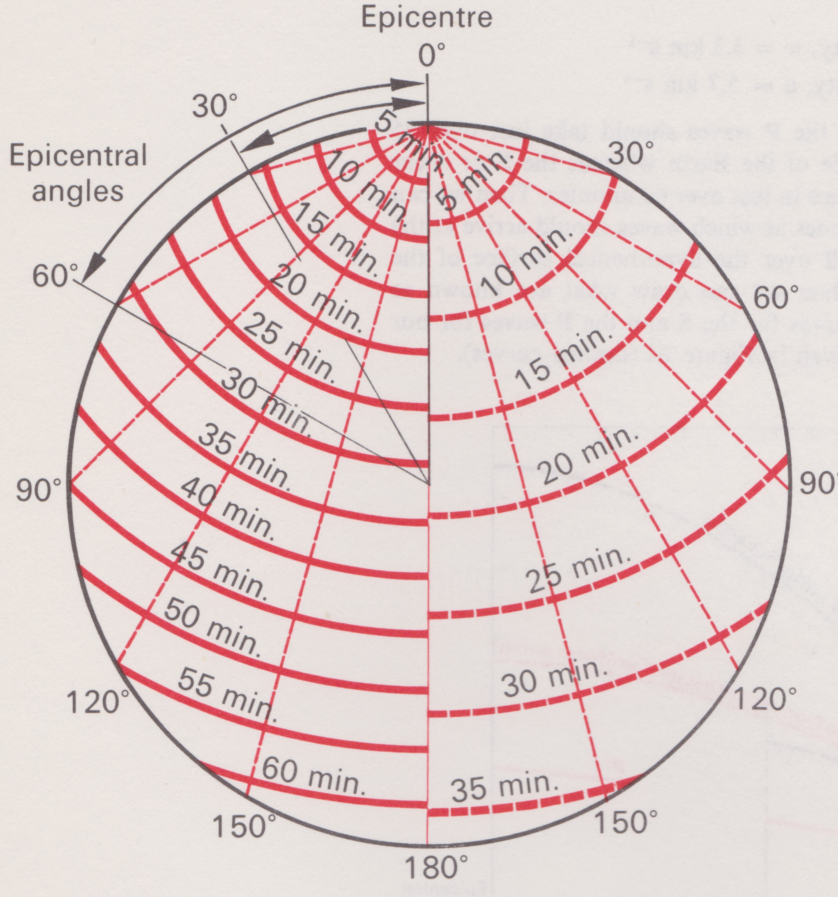


Figure 23 A section through the uniform sphere with the same diameter as Earth showing the travel paths of body waves (dashed lines) to stations at distances 30° apart from the epicentre and the positions of spherical wave fronts (solid curves) at 5 minute time intervals. P waves are shown on the right and have a speed of 5.6 km s^{-1}. S waves, on the left, have a speed of 3.3 km s^{-1}. Travel paths and wave fronts would, of course, extend throughout the sphere.

22.5.2 The simplest Earth model

Suppose we start by imagining the simplest possible model—that the Earth is a uniform sphere, with constant density and elastic properties throughout, and that the density and the elastic constants are the same as we may find in the average piece of rock at the surface of the actual Earth.

Is this a sensible assumption to make? After all, we already know that the average density of the Earth is about 5.5×10^3 kg m^{-3}. As the average density of the rocks at the surface is about 2.9×10^3 kg m^{-3}, it is obvious that some part of the Earth's interior must have a density much higher than the average, and that, therefore, the assumption of constant density is bound to be wrong.

The difficulty is that we also have to assume some values for the elastic moduli, and at this stage of construction of our model we have no idea of what these might be inside the Earth. So, to make a self-consistent set of assumptions as a starting point, we assume that our hypothetical uniform Earth has a constant density of 2.9×10^3 kg m^{-3}, a constant rigidity modulus of 3.16×10^{10} N m^{-2} and a constant axial modulus of 9.40×10^{10} N m^{-2}, these being average values for surface rocks. On this assumption, an earthquake at some point near the surface would send out body waves in all directions, and the wavepaths would be straight lines as depicted in Figure 23.

Furthermore, we can calculate the propagation velocities from the formulae given earlier (24.4.9):

S wave velocity, $w = 3.3$ km s^{-1}

P wave velocity, $u = 5.7$ km s^{-1}

We can see from Figure 23 that the P waves should take just over 35 minutes to arrive at the other side of the Earth whereas the S or shear waves would cover the same distance in just over 60 minutes. Then we can go further and can calculate the times at which waves should arrive at the hypothetical recording stations all over the hypothetical surface of the hypothetical Earth; from these data we can draw what are known as travel-time curves. Travel-time curves for the S and the P waves for our hypothetical uniform Earth are given in Figure 24 (dashed curves).

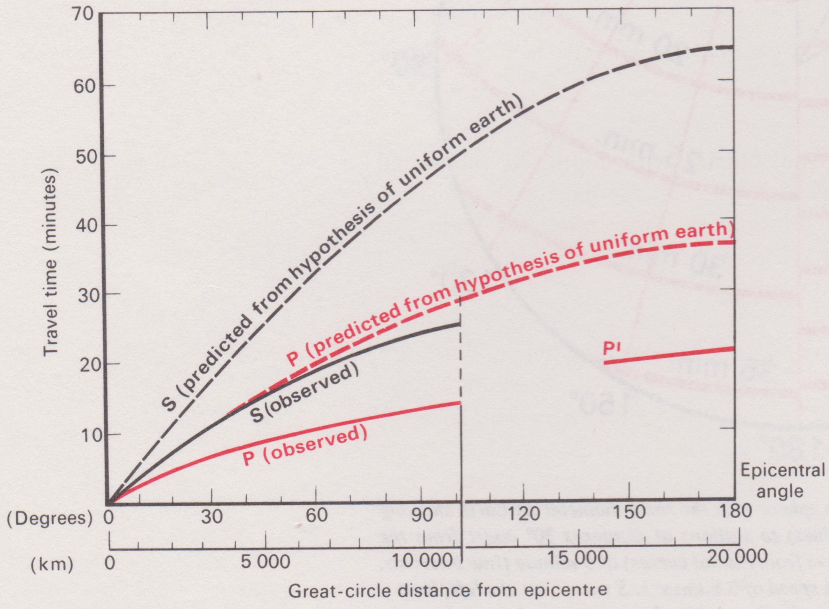


Figure 24 Travel-time curves for body waves in a hypothetical Earth having uniform density (dashed curves), and in the real Earth (continuous curves).

22.5.3 Modifying the simple model

Let us now put on Figure 24 the travel-time curve for an actual earthquake (solid curves). If an earthquake epicentre can be located and its focus is shallow, we can calculate the actual distance to all the recording stations. From seismograms of this earthquake at many stations we can then plot travel-time curves for the seismic waves that make the P and the S deflections.

What can you deduce from Figure 24 regarding the predicted and actual velocities?

As you can see from the figure the arrival times of both the P and the S waves are *earlier* than would be predicted for our uniform homogenous Earth; the divergence between the predicted and the actual times of arrival increases progressively with the distance from the epicentre. Also, beyond a distance represented by 103°, predictions from the hypothetical Earth and actual observations do not resemble each other even remotely; no traces of S waves are recorded and a gap appears in the travel-time curves for the P waves. Obviously, the simple model of a uniform sphere with constant density equal to that of the real Earth's surface rocks and with constant elastic moduli is, as we might have expected, not a good one. It needs to be modified.

How should we modify it?

The first thing to consider is that in reality both S and P waves travel *faster* than is predicted from the uniform Earth model, in which the density and the elastic moduli were assumed to be constant and equal to the values observed in real surface rocks. We worked out a value for the mean density of the Earth right at the beginning of this Unit; and found a value of around 5 500 kg m^{-3}. The observed density of surface rocks varies between 2 500 and 3 000 kg m^{-3}. So there must be much denser material below the surface. Indeed, we would expect the density to increase with depth, due to compression of the underlying material by the weight of the overlying rocks. Alternatively, the density increase could be due to a change in chemical composition—that is, to the presence of some intrinsically denser substance, such as a heavy metal, in the Earth's interior.

So the density in the Earth must increase with depth; but let us assume for the moment that ψ and μ do not change with depth. As the actual densities in the Earth are *higher* than the density we have assumed, then the actual velocities should be *lower* than the ones predicted from our model, instead of the other way about, because the velocities of both P and S waves are inversely proportional to the square root of density (see equations (6) and (7)). Furthermore, as the angular distance between epicentre and recording station increases, the travel paths of the body waves going directly from epicentre to station penetrate more and more deeply into the body of the Earth, where they pass through material of greater and greater density. So, the difference between the actual velocities and the predicted ones should get bigger and bigger. We should thus expect a situation something like that in Figure 25.

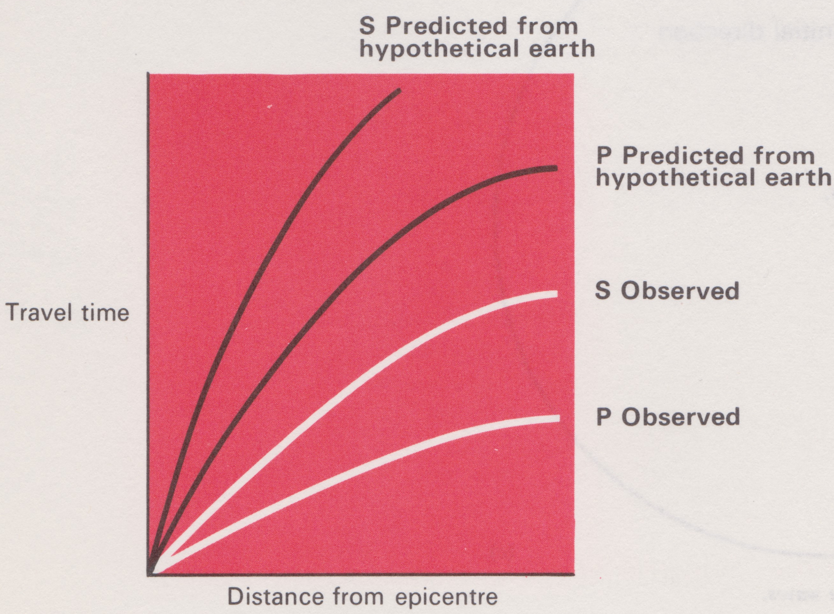


Figure 25 Travel-time curves as observed (continuous curves), and as predicted from a hypothetical Earth in which density increases with depth but ψ and μ remain constant (dashed curves).

Evidently the assumption that the elastic moduli are the same as for surface rocks and that they do not increase with increasing depth, *must be wrong*. To explain the discrepancy between the observed and predicted velocities, at least up to the 103°, we *must conclude* that both the axial modulus and the rigidity modulus increase with depth more rapidly than does the density. We then have the picture of waves being propagated downwards into a medium in which their velocity increases with depth.

What will happen to them?

If you refer back to section 22.4.11 you will see that if a wave passes from a medium in which it travels at a given speed into a medium in which it travels faster, then it will be refracted away from the normal to the surface in the way shown in Figure 26 (a). The angle of refraction will therefore be greater than the angle of incidence, and both angles will, of course, get progressively larger if several such boundaries are encountered. If, instead of increasing in jumps as shown in Figure 26 (a), the velocity increased continuously with depth, then it is easy to see that the ray would

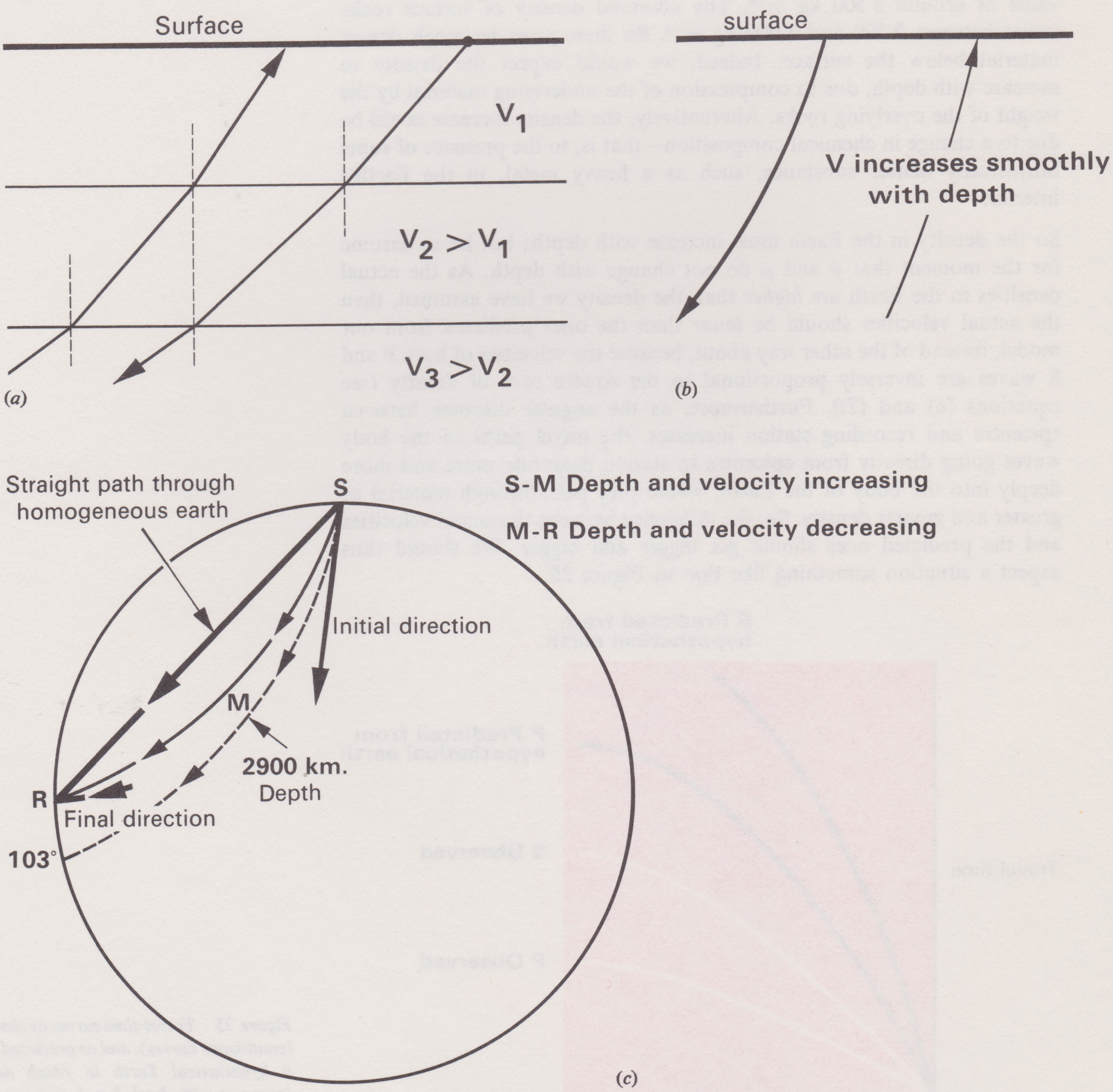


Figure 26 The refraction of earthquake waves.

be curved, as shown in Figure 26 (b). There will come a time when the angle of incidence is so high (greater than the critical angle) that the waves will be totally reflected. They now come *out* towards the surface, going through a medium in which the velocity is progressively *decreasing*. So, the angle of incidence and refraction will now get progressively smaller. The outcoming ray will be curved in just the same way as the ingoing one (Fig. 26 (c)). We would therefore expect a seismic wave starting from a point such as S (Fig. 26 (c)) to follow a curved path, convex with respect to the Earth's centre.

Our modified model is then no longer that of a *uniform* Earth, but of a *continuous* one with no sudden breaks or discontinuities in its structure.

By making appropriate assumptions about the rates at which ρ, μ and ψ increase with depth, we can get our predicted travel-time curves to fit the observed ones—until everything goes wrong at an epicentral angle of 103°. It seems that waves passing through the Earth at depths down to a particular limit behave in the way we have described, but that at greater depths than this there must be some sudden drastic change in the properties of the Earth's interior. Having specified how ρ, μ and ψ vary with depth, to make the model fit the observations, one can calculate the corresponding wave paths, and hence the depth at which the change occurs. The mathematics is complicated and of no special interest to us. The result obtained is that body waves passing through stations just inside the 103° limit penetrate the Earth to a maximum depth of 2 900 km (Fig. 26 (c)).

However, three features of the travel-time curves in Figure 24 still remain to be interpreted:

(i) no direct S waves are recorded at distances from the epicentre corresponding to angles greater than 103°;

(ii) between 103° and 143° no direct body waves of any kind are received;

(iii) between 143° and 180° direct P waves are once more received, but their travel time is longer than would be expected from an extrapolation of the travel-time curve beyond 103°.

Why do we use the term *direct* waves in (i) to (iii)?

Because, between 103° and 143° earthquake waves *are* received, but these have usually been reflected from the Earth's surface as shown in Figure 27. They can be identified as reflected waves as they have longer travel paths and therefore take longer to arrive at the recording station.

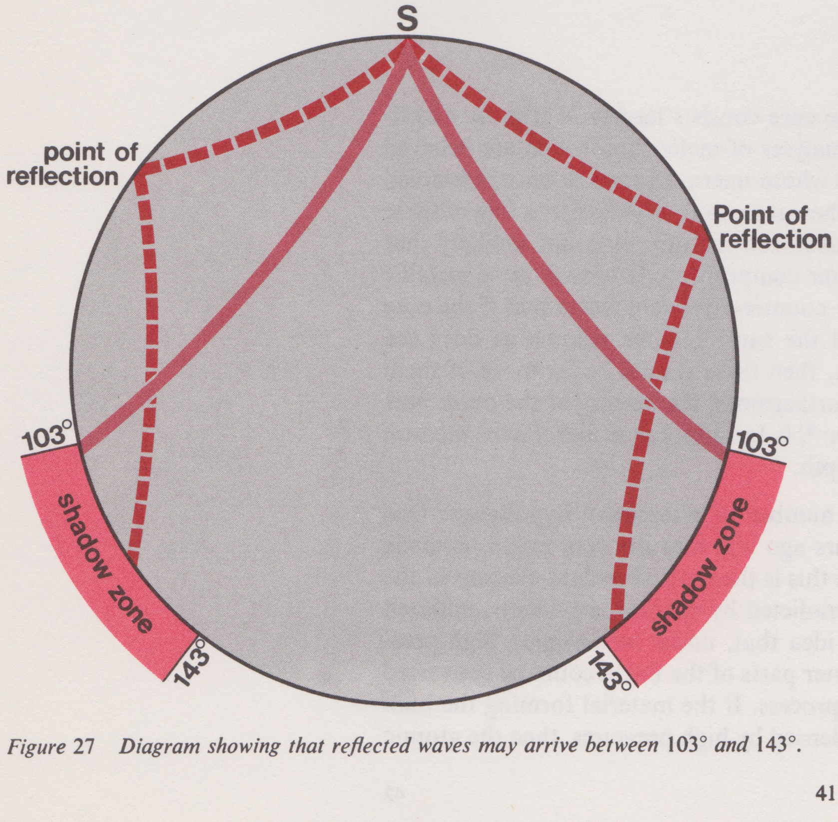


Figure 27 Diagram showing that reflected waves may arrive between 103° and 143°.

22.5.4 The centre of the Earth at depths greater than 2 900 km

There is evidently a sudden, qualitative change—a *discontinuity*—in the composition or structure of the Earth at this depth. The boundary surface at which this sudden change occurs is called the *core-mantle boundary*. In the part of the Earth outside this boundary—the *mantle*—both S and P waves are propagated. The velocity of propagation here increases with depth, as do the density and the elastic moduli—though the elastic moduli increase more rapidly than density. But what happens across the core-mantle boundary, in the *Earth's core*? S waves are no longer propagated. P waves are bent (refracted) in some odd way, so that they only reappear at epicentral angles greater than 143°, and when they do reappear they arrive later than expected.

discontinuity

core-mantle boundary

Perhaps the most significant of these facts is that S waves stop at the core-mantle boundary.

What does this suggest to you about the possible composition of the core?

Refer back to section 22.4.9 if you have difficulty in answering this question.

Shear waves cannot be transmitted through a liquid. So it seems reasonable to infer that the core—or at least the outer part of it—has the properties of a liquid. The use of the terms 'solid' and 'liquid' in connection with the huge pressures that prevail in the Earth's interior is questionable. What we mean by the term 'solid' in this context is simply that the elastic behaviour of the materials in question can be described by equations which match those applying to ordinary solids in normal laboratory conditions. These equations involve the use of two coefficients, 'compressibility' which is the measure of resistance to pressure and 'rigidity' signifying a resistance to shearing stress. In the case of a liquid, the resistance to shear is effectively zero. This is why a liquid does not transmit S waves.

We can now infer that all the Earth outside the core is solid (except possible for minor pockets), because both S and P waves travel through it, and that at least the outer part of the Earth's core has the properties of a liquid, solely because it will not support S waves.

22.5.5 Composition of the core

It has long been assumed that the core consists largely of iron or nickel-iron. This view is supported by analyses of meteorites, which are believed to be pieces of an exploded planet whose internal structure once resembled that of the Earth. The argument then goes on to say that, as a few of these meteorites are composed of a nickel-iron mixture, it is not unlikely that the centre of the Earth is of the same composition. However, these metallic meteorites are rather rare and the counter-argument states that if the core of this exploded planet occupied the same relative volume as does the Earth's core relative to the Earth, then there should be far more of them landing on the Earth's surface. Furthermore, the density of the outer part of the Earth's core, 9.7×10^3 kg m^{-3}, is too low for a nickel-iron mixture at the pressures present at this depth.

These arguments have led to a number of alternative hypotheses. One suggestion made about thirty years ago was that the core might be made up of compressed hydrogen, since this is the most abundant element in the Universe. This theory, while contradicted by weighty arguments, initiated new investigations based on the idea that, under increasingly high pressures, the minerals forming the outer parts of the Earth could be converted to higher-density forms by some process. If the material forming the bulk of the Earth were sufficiently condensed by high pressures, then the atomic

structure would be crushed like an egg-shell and the normal pattern of a central nucleus with surrounding electrons would be deformed. Material in this state would be of the right density for the Earth's liquid outer core and also, as there is no 'rigid' atomic lattice to be maintained, it would have the properties of a liquid, explaining the fact that S waves cannot travel through it.

All this is saying is that we really don't know what the Earth's core consists of, although we know that it is dense and that it behaves as a liquid. In Unit 23 we shall see that it also must have properties which can support electric currents and generate magnetic fields. Any substance which has been deformed in the way we have described will meet these requirements, whether it be hydrogen, helium, nickel and iron, or silicon. Silicon compounds are a strong contender, if only by virtue of the fact that they are the most abundant minerals in the rest of the Earth.

22.5.6 The inner core

Recent investigations into the structure of the Earth's core have revealed that it is not liquid throughout; the P waves travel with slightly increased velocity through its central part which indicates that the very centre of the Earth must be 'solid' (see Fig. 24). The interpretation here, on the theory of deformation and compression, is that when pressures reach a value in the order of 3 million atmospheres, as they must at the centre of the core, then even the deformed material has to become rigid, being squeezed together as a solid mass, and therefore behaves as a solid through which P waves can travel more quickly.

Because there is a sudden increase in density at the core-mantle boundary, P waves must be refracted there. Look at Figure 28 and you will see that a P wave (A–B) which just impinges on the margin of the core will be strongly refracted towards the Earth's centre, because the angle of incidence is large and the wave is entering a medium where it travels more slowly. The path through the outer core is convex inwards because the waves travel more quickly with increased depth of penetration. Then, on leaving the core at C, it is once more refracted on entering more rigid material. Finally, it will reach the Earth's surface at D, almost opposite the epicentre.

Note that the angles of incidence and refraction must satisfy Snell's Law (equation (11) in section 22.4.11). The velocities of P waves on either side of the core-mantle discontinuity are known, from analysis of data of the kind we are discussing, to be 13.64 km s^{-1} just outside the discontinuity and 8.10 km s^{-1} just inside it (see Table 2).

A P wave such as AE will hit the core at a smaller angle of incidence, will therefore be refracted less on entering and leaving the core, and so come back to the surface nearer to the epicentre than in the case of the core-brushing wave. The smallest epicentral angle such a wave can have is 143° (see Fig. 24) and it will follow a path approximating to AEFG of Figure 28. Remember that a direct P or S wave that brushes past the core, without touching it, reached the surface 103° from the epicentre. S waves cannot pass through the core, so over the whole of the area between the two 103° and 143° no direct waves of any kind are recorded, and between the two 143° marks only direct P waves emerge.

Here is a simple experiment you can do to illustrate the effect we have been describing. All you need is a transparent cylinder that can be filled with water. The measuring cylinder in your home experiment kit will do nicely.

Take the measuring cylinder, fill it with water and hold it vertically in front of a table lamp. If a piece of plain paper is placed on the opposite side of the cylinder to the lamp, you will see on it a bright central zone bounded by darker zone on either side. These dark zones are caused by the refraction of the light by the water in the cylinder and are analogous to the seismic shadow zones between 103° and 143°. If the water had been in a sphere instead of a cylinder the analogy would be even closer for then the 'dark' shadow zone would be a ring instead of two lines. Try a goldfish bowl—no need to remove the fish!

The P waves just described passed only through the outer part of the core. If a wave hits the core more head on, then it will not be refracted so much and will pass right through the centre of the core. It has been found that travel-times for such waves are less than they should be if the entire core were liquid. This is why it is suggested that the very centre of the Earth, the *inner core*, is solid. Once again using the discrepancy between estimated and actual travel-times, the size of the solid inner core (diameter 3 500 km) can be deduced. The simplest case is, of course, the wave that travels vertically downwards from the epicentre and is not refracted at all, such a wave is shown as line AHKL in Figure 28.

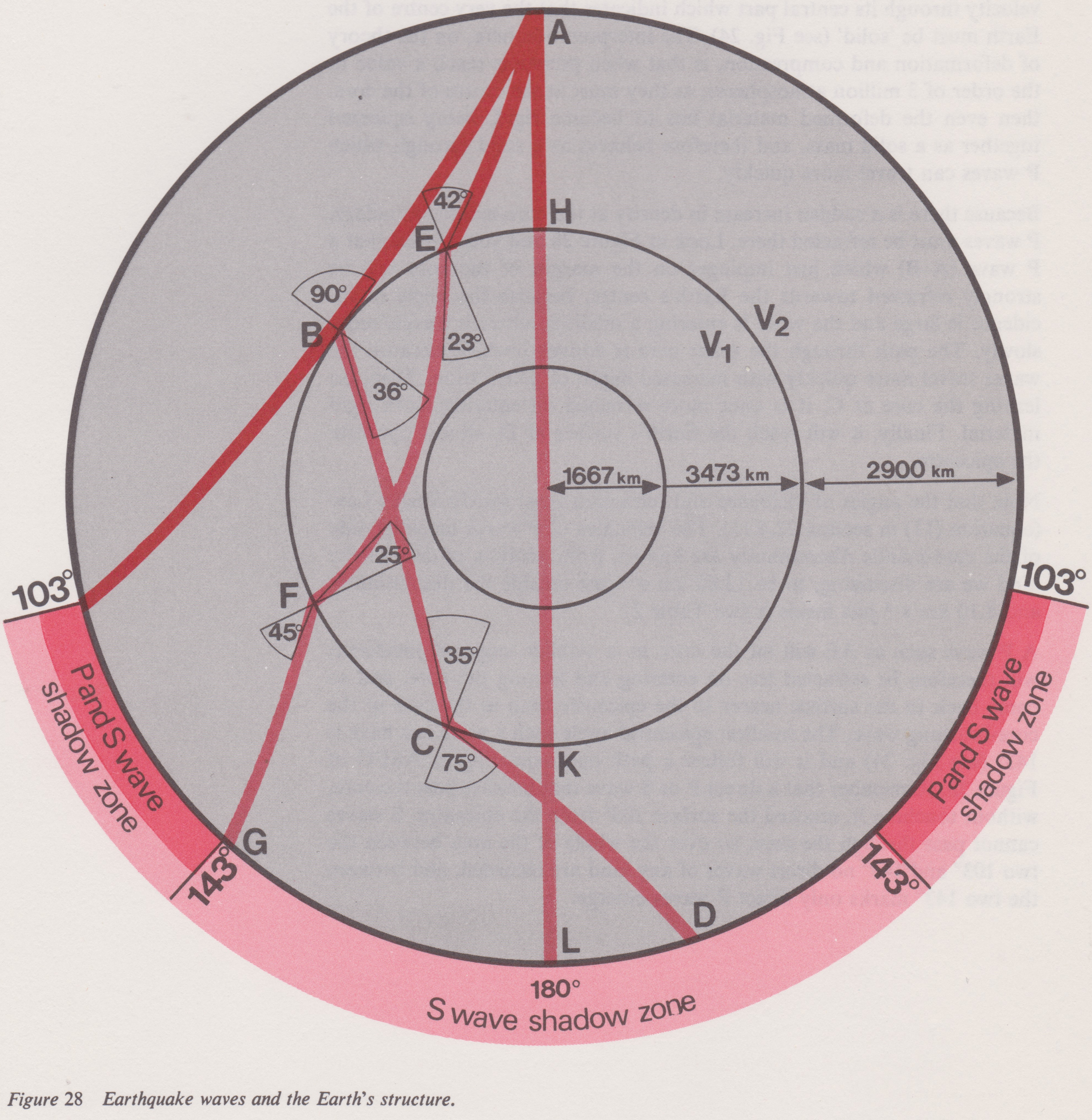


Figure 28 Earthquake waves and the Earth's structure.

22.5.7 The interior of the Earth near the surface

From the Earth's deep interior let us now turn to the outer part of the Earth, that region within 100 km of the surface but still beyond man's direct observations. Early in the twentieth century, a Yugoslav seismologist, Mohorovičić, noted that two distinct sets of P and S wave deflections appeared on seismographs when an earthquake within a radius of 800 km and with a focus within 40 km of the surface was recorded.

Bearing in mind that waves are refracted within the Earth, how do you think this could happen?

Mohorovičić concluded that the only way this could happen was if waves travelled from the focus to the recording stations by different paths. The first set of waves travelled, he suggested, directly through the Earth whereas the second set took longer to arrive because they had been refracted. The most reasonable postulate was that this second set of waves had initially travelled downwards from the focus (Fig. 29), then, on encountering a zone where they would travel faster, had been refracted, only to be refracted again towards the surface when they re-encountered this boundary because of the Earth's curvature.

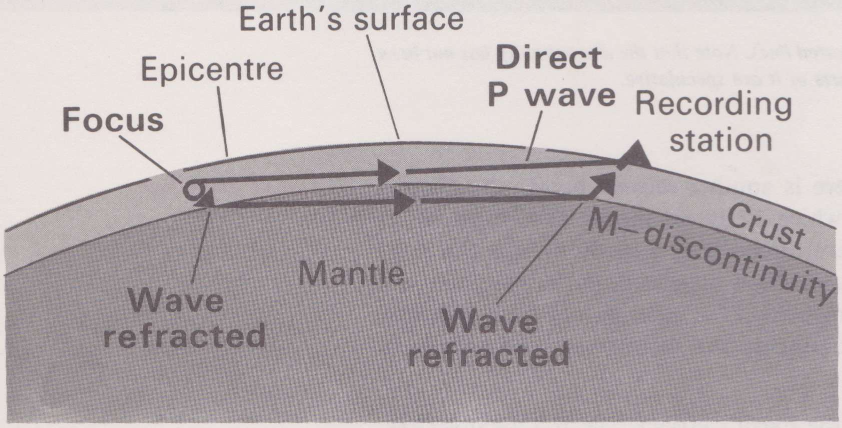


Figure 29 The Mohorovičić discontinuity.

Mohorovičić inferred that within 100 km of the surface of the Earth there were two zones—an upper one where seismic waves were propagated slowly and a lower one where they travelled more quickly. The surface between these zones was thus a 'seismic discontinuity'. Later, when it had been found that this discontinuity was present around most of the Earth, it was named after Mohorovičić and is now known as the *Mohorovičić* or *M* discontinuity or *Moho* for short. (It should not be confused with the 'Mohole', the name given by the Americans to a now abandoned project to drill a hole through the Mohorovičić discontinuity and find out what lay underneath.)

Mohorovičić discontinuity

The Mohorovičić discontinuity is now accepted as a major dividing line in the structure of the Earth, the part above it being termed the *crust* and the part below, as far down as the core boundary, being known as the Earth's *mantle*. We therefore have three major zones in the Earth. From the surface down they are: the *crust*, separated by the Mohorovičić discontinuity from the *mantle*, which in turn is separated by the core-mantle boundary from the *core*.

22.5.8 The Earth's crust

The depth of the Mohorovičić discontinuity, and thereby the thickness of the Earth's crust, has now been determined in most places around the Earth. Under the ocean basins the crust is thin, averaging 5 km in most places, whereas under the continents it varies between 20 and 65 km. In the continental regions under the extensive flat areas such as the Prairies of the American mid-west and the Steppes of Russia, the depth to the Moho is between 30 and 35 km. However, under the major mountain ranges like the Alps, the Himalayas and the Cordilleras of the western part of South America, it thickens to 60–65 km. It can be seen in Figure 30 that the configuration of the Moho mirrors, in an exaggerated way, the topography of the Earth's surface.

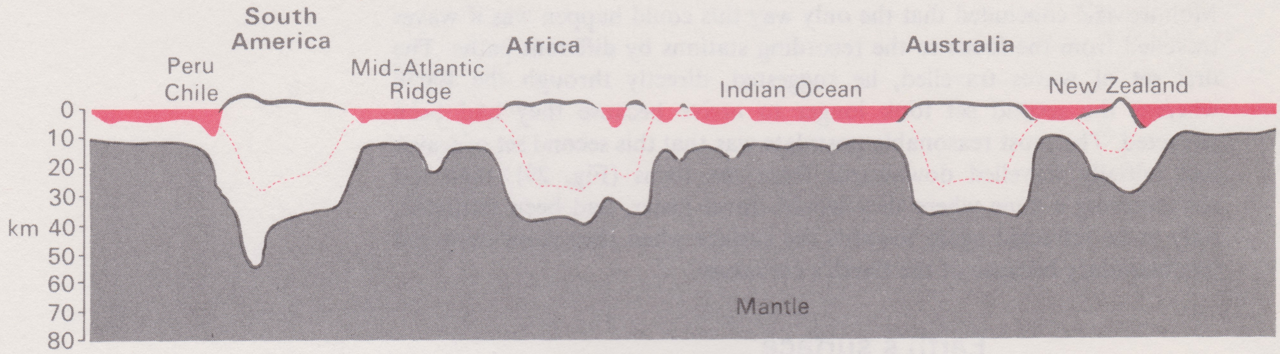


Figure 30 *The Conrad discontinuity (dotted line). Note that the discontinuity has not been detected along the whole dotted line; parts of it are speculative.*

Within the continental crust there is another seismic break, the *Conrad* seismic discontinuity (Fig. 30) which is by no means as distinct or as continuous as the Moho but nevertheless conveniently divides the crust into an upper and lower part. It has been suggested, in the past, that the thin oceanic crust may be the equivalent of the lower part of the continents. But it is the upper part of the continents that can be seen and examined. So let us consider this first.

Conrad discontinuity

Even a brief look at the rocks of the Earth's surface reveals a great variety. There are rocks formed by crystallization from molten rock material such as the lavas from volcanoes; there are rocks which are ancient sediments, comparable with the present-day sediments we see on beaches and in river banks, but many millions of years older than them; finally, there is a wide variety of rocks that started out as one type but have since been vastly altered by being reheated or deformed by pressures and stresses within the Earth.

However, despite this wide variety, two elements, silicon (Si) and aluminum (Al) predominate in most surface rocks. The collective term that is used for such rocks combines the chemical symbols for these two most abundant elements (Si and Al) to give the word *sial*. The upper part of the Earth's continental crust is commonly called the Sial or the sialic layer. In more detail it is considered to have the composition of a variety of granite†, granite being the most abundant rock type in this upper part of the Earth's crust.

Sial

The lower part of the Earth's crust presents more of a problem. The Sial is not present under the ocean basins where the crust is, on average, only 5 km thick. Most of the oceanic crust lies below 3 to 5 km of ocean, and the only land masses are isolated islands. It is nevertheless striking that all these islands, lying within the deep ocean basins, are of volcanic origin and are formed of a rock called basalt†. Furthermore, during the last two

decades, many rocks have been dredged from the deep ocean floor and, once again, excepting a thin veneer of sediment, these are mostly basalts. A basalt resembles granite in that it consists mainly of silicon and aluminium. It differs from granite in having a high magnesium content, which granite lacks. For this reason, and purely to get a convenient descriptive label, the oceanic crust is commonly called the *Sima*: Si for silica and Ma for magnesium. (Note: Mg is the actual symbol for Magnesium—see Unit 6.) **Sima**

What then constitutes the lower part of the continental crust? Although from Figure 30 it may appear to be a continuation of the oceanic crust, it is probably not the equivalent of the oceanic crust. But there is not enough evidence to make any definite alternative statement. What we can say at this stage is that: (1) the oceanic and continental crusts are not alike; (2) the simatic oceanic crust is probably basalt; (3) the continental crust has two parts, (a) an upper sialic zone consisting of a heterogeneous collection of rocks of which granites are the most abundant part; and (b) a lower part differing from the sialic upper part and also from the simatic oceanic crust.

One thing, however, is common to both the continental and the oceanic crusts. They are both, by definition, separated from the underlying mantle by the Mohorovičić discontinuity.

22.5.9 Composition of the upper mantle

The sudden change in seismic velocities at the Moho (P waves increase from 6.9 to 8.1 km s^{-1} and S waves from 3.9 to 4.6 km s^{-1}) must be due *either* to a difference in chemical composition, or to a phase change.* Accepting, as most Earth scientists do, that the rocks above the Moho are basalts or, more probably, their coarse-grained equivalents known as gabbros,† then, if the Moho represents a chemical change, the upper mantle must be composed of rocks whose minerals are denser and more rigid than those of gabbros. At the temperatures and pressures which must be present in the upper mantle, the most likely contender to give the appropriate seismic velocities is a rock called *peridotite*†. Peridotite has a density of 3.3×10^3 kg m^{-3} and consists almost entirely of the magnesium and iron silicates, olivine $(Mg, Fe)_2SiO_4$, and pyroxene $(Mg, Fe)SiO_3$.

From the velocities at which seismic waves travel through rocks, an equally viable contender for the upper mantle is a rock called *eclogite*. Eclogite has the same composition as a basalt, or for that matter a gabbro†, but the atoms of its minerals are more densely packed.

On seismic data alone, peridotite and eclogite are equally acceptable. However, various other considerations make peridotite the only possibility for the upper mantle under the oceans and also, though with less certainty, for the sub-continental upper mantle. The prime factor here is that pressures equivalent to a depth of about 60 km are needed before the dense minerals of eclogite will form. The Moho under the oceans is only 11 km down and only under the very deepest parts of the continent is this requirement for the formation of eclogite met. Secondly, some volcanoes, whose lavas were probably generated about 60–100 km down, bring to the surface solid pieces of the upper mantle torn from the walls of the volcanic conduit by the ascending lava; in oceanic regions these fragments of the

* *A phase change, in this sense, means a change in the physical properties of a substance without* any *change in composition. For instance both graphite and diamond are* phases *of pure carbon. Diamonds are formed at high pressures deep in the Earth; the atoms are tightly packed together so the diamond is hard and dense. Graphite is formed at low pressures; the atoms are loosely packed and so graphite is softer and less dense.*

mantle are almost invariably peridotite. Thirdly, there are several places where, through violent Earth processes, parts of the upper mantle have been brought close to the surface and subsequently laid bare by erosion for our inspection. In all these cases, peridotite is the dominant rock type present. Fourthly, as with the core, meteorite evidence is valuable, for as well as the metallic meteorites representing, supposedly, the core of an exploded planet, there are stony meteorites of peridotitic composition which, it is argued, represent the mantle of the planet.

As well as these four lines of evidence there are numerous more sophisticated arguments, accepted by most Earth scientists, the majority of which support the contention that the upper mantle, with the possible exception of that under the thickest parts of the crust, is peridotite.

22.5.10 The deeper parts of the mantle

The mantle, which forms 80 per cent of the Earth's volume, is probably composed throughout of magnesium and iron silicates. Physical data show that from the top to the bottom, from just below the Moho down to the core-mantle boundary at 2 900 km, the velocity of the P waves increased from 8.1 to almost 14 km s^{-1} (Table 2). The density increases from 3.3 to 5.6×10^3 kg m^{-3} and the pressure from 9 to almost 1 400 kilobars (a kilobar is approximately equivalent to 1 000 times atmospheric pressure). There are two seismic discontinuities within the mantle at about 400 and 700 km and these probably represent phase changes, where the main mineral components of the mantle are forced, by ever increasing pressure related to depth, to readjust their crystal lattices so that the atoms are more densely packed.

22.5.11 The low-velocity layer

Seismic velocities increase with depth throughout most of the mantle. But, as early as 1926, Gutenberg noted that seismic waves from earthquakes with foci at depths of between 50 and 250 km took slightly longer to arrive at the recording stations than they should on theoretical grounds. He therefore postulated that between these depths there was a low-velocity layer or channel. Subsequently, and particularly since man-made explosions have been used as well as naturally-occurring earthquakes, the existence of this low-velocity layer has been confirmed. In most respects it has the same configuration as the Moho, being deeper under the continents than it is under the oceans. There is still some doubt, however, as to whether it represents a gradual diminution of wave velocities over the whole zone from about 50 to 250 km or several very narrow layers within the zone where the seismic velocities are more markedly slower. Figure 21 illustrates that waves take longer to pass through a low-velocity channel, not only because of their low velocity but also because they follow a zig-zag path.

low-velocity layer

The low-velocity layer represents a reduction of P wave velocities from 8.1 km s^{-1} at just below the Moho to 7.8 km s^{-1} at about 100 km depth, rising again to 8.1 km s^{-1} at somewhere around 250 km. Over the same distances the S wave velocity changes from 4.6 km s^{-1} to 4.4 km s^{-1} and back to 4.6 km s^{-1}. The variation of S and P wave velocities with depth obtained by two investigators are shown in Figure 31. There is at least a qualitative agreement, and both sets of results show a velocity minimum somewhere between 70 and 150 km depth. The presence of the low-velocity layer is well proven; and the composition of the upper mantle is generally accepted as peridotite. How can there be a low-velocity layer at all? One of the currently favoured explanations depends on an effect of temperature. For the time being we shall just assume it to be a fact that the temperature of the Earth increases with depth. A curve plotting

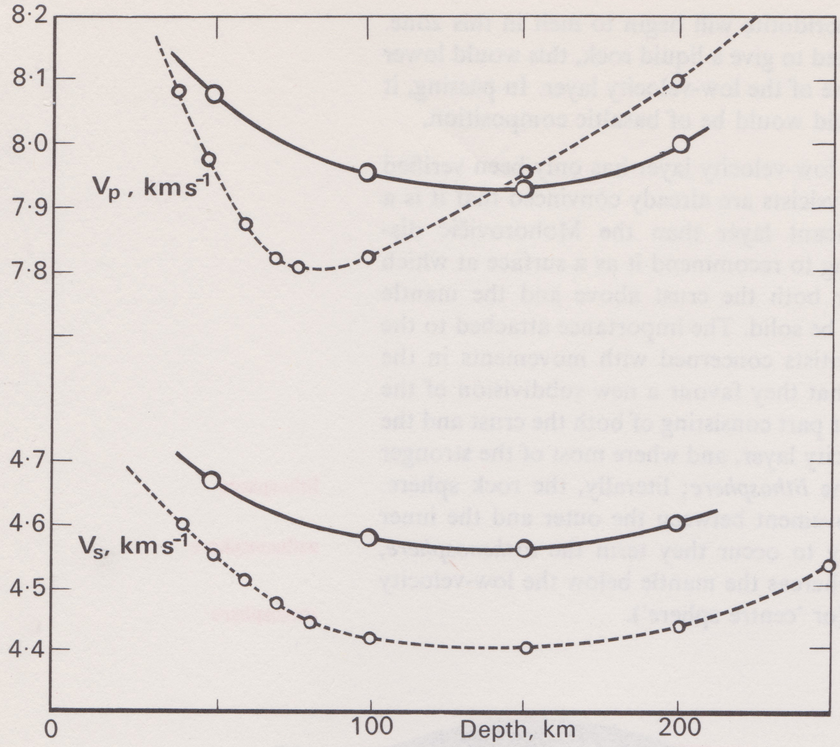


Figure 31 Experimentally-determined variation in P and S wave velocities with depth.

temperature against depth is known as the Earth's *thermal gradient* (Fig. 32). We can also get from experimental data the temperature at which peridotite will begin to melt for a range of pressures corresponding to varying depths. When this curve, known as the *incipient melting-point curve* for peridotite, is also plotted on Figure 32, it can be seen that the two curves approach each other at pressures equivalent to depths of between 50 and 250 km. Now, the position of the melting-point curve is by no means fixed. If there is water in the peridotite, even as little as 0.1 per cent, then the melting-point curve is lowered appreciably (dashed line on Fig. 32) and will intersect with the thermal gradient curve between X and X′ on

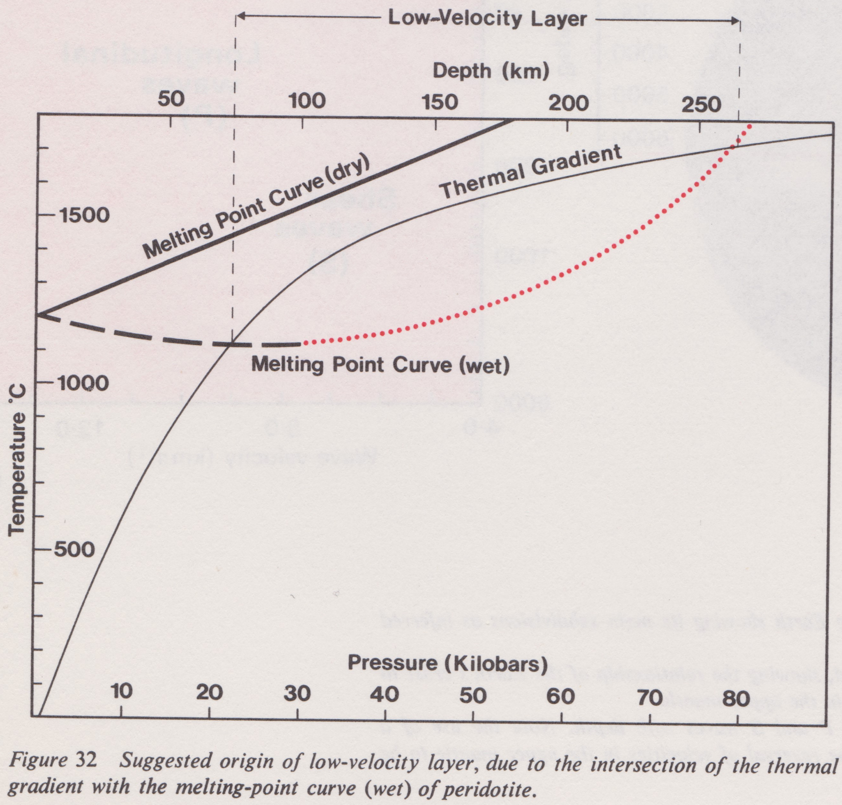


Figure 32 Suggested origin of low-velocity layer, due to the intersection of the thermal gradient with the melting-point curve (wet) of peridotite.

Figure 32. This means that the peridotite will begin to melt in this zone. If 5 per cent of the peridotite melted to give a liquid rock, this would lower the seismic wave velocities to those of the low-velocity layer. In passing, it is interesting to note that the liquid would be of basaltic composition.

Even though the existence of the low-velocity layer has only been verified in the last few years, most geophysicists are already convinced that it is a far more important and significant layer than the Mohorovičić discontinuity. The Moho has nothing to recommend it as a surface at which movement could take place, for both the crust above and the mantle below must, on seismic evidence, be solid. The importance attached to the low-velocity layer by Earth scientists concerned with movements in the outer part of the Earth is such that they favour a new subdivision of the Earth. They suggest that the outer part consisting of both the crust and the upper mantle above the low-velocity layer, and where most of the stronger earthquakes occur, be termed the *lithosphere*; literally, the rock sphere. The low-velocity zone where movement between the outer and the inner parts of the Earth is most likely to occur they term the *asthenosphere*, meaning literally weak sphere, whereas the mantle below the low-velocity layer is labelled the *mesosphere* (or 'centre sphere').

lithosphere

asthenosphere

mesosphere

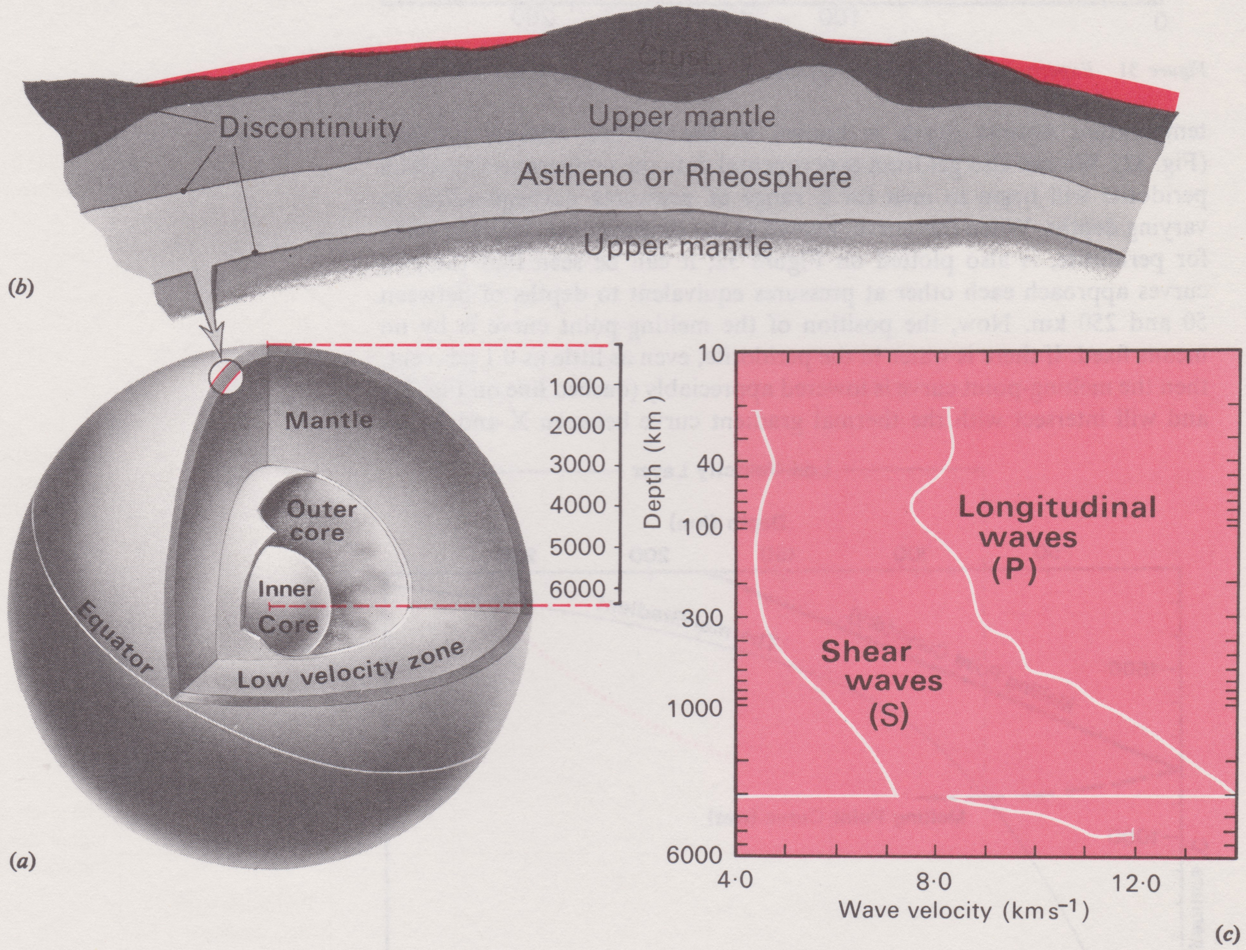


Figure 33 (a) Cut-away section of the Earth showing its main subdivisions as inferred from seismic data.

(b) Enlargement of segment indicated, showing the relationship of the Earth's crust to surface features and the asthenosphere in the upper mantle.

(c) The variation in the velocity of P and S waves with depth. Note the use of a logarithmic depth scale which allows the reversal of velocities in the upper mantle to be depicted.

Figure 33, a cut-away section of the Earth, shows the internal structure as deduced from seismic evidence. In Table 2, the dimensions and other physical facts concerning the various zones are listed. We do not suggest that you commit all these data to memory, but rather that you use them for reference purposes.

Finally, Figure 34 shows how the density of the Earth varies with depth in diagrammatic form.

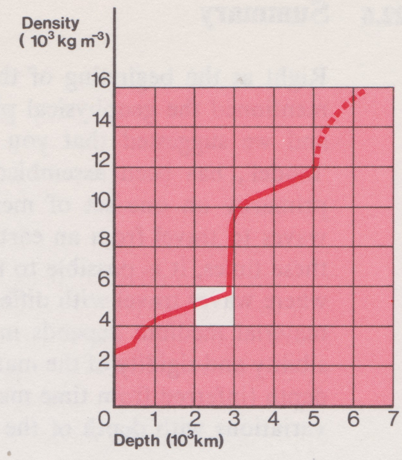


Figure 34 Density variation in the Earth's interior.

Table 2 Properties of the Earth's Layers

Zones and discontinuities	Depth to boundaries (km)	Velocity of P waves (km s^{-1})	Percentage of volume	Density (10^3 kg m^{-3})	Pressure (kilobars)	Nature of regions	Possible nature of boundaries
CRUST (Sial)	Continental average, 33	6.5		2.7		Heterogeneous Solid	
Conrad — — — —	Mountain ranges, 65		1.55	2.8		Oceans basaltic Continents granitic	
CRUST (Sima)	Oceanic (from sea level), 10–11						
		6.9		2.9	9		Chemical change from basalt above to peridotite below
—Mohorovičić—				3.3			
UPPER MANTLE		8.1				Various types of peridotite	
— — — —	— c.50 —						
Low-velocity layer		7.8				Partly fused peridotite	
— — — —	— c.250 —						
UPPER MANTLE		8.1	82.25			Peridotite with high density minerals	Phase change to higher density minerals
	— c.1 000 —						
LOWER MANTLE		10.7		4.3	270	Peridotite with high density minerals	
		13.6		5.7	1 368		Change from silicate mantle to metallic core
— Core-mantle —	2 900						
		8.1		9.7			
OUTER CORE		10.3		11.8	3 180	Probably iron-alloy nickel Liquid	Phase change from liquid to solid
	— 5 000 —						
INNER CORE		11.2	16.20	c.14	3 300		Transition zone of about 100 km
CENTRE	6 371			c.16	3 600	Iron-nickel Solid	

22.6 Summary

Right at the beginning of this Unit, when we were stating its aims, we mentioned the geophysical picture of the internal structure of the Earth, and we suggested that you took careful note of how this picture, or 'model', has been assembled. You have seen that the model is based primarily on one set of measurements—of the times taken by seismic waves to travel from an earthquake to various recording stations. From these times, it is possible to infer the existence of layers within the Earth where waves travel with different velocities. But the velocity of waves in a material medium depends in a particular way on the density, compressibility and rigidity of the material. So the variations in wave velocities with depth (inferred from time measurements) could be interpreted in terms of variations with depth of the physical properties of the rocks.

Starting with the simplest model (a uniform Earth), we introduced refinements, step by step, to account for the observed data. This procedure is quite a common one in science. It is useful as long as the model does not have to be changed to 'fit' every new piece of information that comes along, thereby becoming more and more complicated. This happens quite often too, and usually means that a new basic scientific principle is needed, on which to found a new and simpler model.

This structural model of the Earth's interior seems to be quite a good one, but there is evidently much room for speculation—and for further investigation—about the composition of the materials that could have the right physical properties. For example, peridotite is the best candidate for the upper mantle, but no one has ever seen it in place—nor is anyone likely to. On the seismic information Earth scientists have at present, the upper mantle could just as well be green cheese, as long as it is green cheese with a density of about 3.3×10^3 kg m^{-3} and the correct elastic moduli.

Reading List

The following articles in *Understanding the Earth* relate to material in this Unit:

Chapter 3	The composition of the Earth.	P. G. Harris
Chapter 6	The Chandler wobble.	M. Chinnery
Chapter 7	The Earth-Moon system.	Z. Kopal
Chapter 8	Meteorites.	B. Mason
Chapter 23	Earthquake prediction and modification.	R. L. Kovach
Chapter 24	Nuclear explosions and earthquakes.	D. Davies

The rotation of the Earth—some further consequences

In section 22.3.1 we pointed out that photographic evidence shows the Earth, the Moon and the planets to be more or less spherical. But if someone were to ask you: 'why is this so? Is it just a coincidence?'—how would you answer? If it were said that one of the planets, Pluto for instance, is shaped like a banana or a beer barrel, would you believe it? And if not, why not?

Refer back to Unit 4 and the law of gravitation. If only the law of gravitation were involved, the Earth and planets would be exactly spherical because that would be the configuration in which all bits of matter in these bodies could get as near as possible to their respective centres.

In fact, as we said in section 22.3.1, the shape of the Earth is an oblate spheroid. But why? It may have occurred to you already that this is a consequence of the Earth's rotation, since the centripetal force would make the Earth bulge at the equator and, consequently, flatten at the poles.

But this is only one of the consequences of the Earth's rotation. Another, and an important one, is that all things moving over the surface of the Earth—such as rockets, shells, aeroplanes and, on a grander scale, the atmosphere and waters of the oceans—tend to drift sideways: to the right in the northern hemisphere and to the left in the southern hemisphere. This drift is the *Coriolis Effect*.

Coriolis Effect

Let us consider what happens when a rocket (Rocket I, see Fig. 35) is orbiting around the Earth at a constant speed and a fixed distance from the Earth's centre. The rocket's orbit is in a plane containing the axis of rotation of the Earth and fixed with respect to the stars. (There is no net force on the rocket. The centrifugal force due to its curved trajectory is exactly counter-balanced by the gravitational force attracting it towards the centre of mass of the Earth.)

The rocket is being observed by four observers on the surface of the Earth, all of them sitting on the same meridian. It does not matter which meridian we choose. Let us choose the Greenwich or zero meridian. One observer (A) is at the North Pole (90° latitude), another (B) is at 60°, the third (C) is at 30° and the fourth (D) is on the equator (0°). Let us suppose that the four observers start observing the rocket just at the moment when the Greenwich meridian lies in the plane of the rocket's orbit. Let us suppose, further, that just at that moment the rocket is immediately above the North Pole, that is on the point P_0 on the Earth's axis of rotation, as shown in Figure 35. And let us assume, finally, that the rocket's speed and distance from the Earth's centre is such that it makes one orbit round the Earth every 10 hours. This last assumption means that the rocket will take just 3 000 seconds to move one-twelfth of a turn—from P_0 to P_1 (or from P_1 to P_2 or from P_2 to P_3).

By the time the rocket is at P_1, where is observer B? Due to the Earth's rotation, B will now be at B′ (Fig. 36). You can easily calculate the angle BEB′—it is 12.5 degrees. 3 000 seconds later, when the rocket is at P_2, C will have moved round to C′—that is, through the angle CFC = 25 degrees; and 3 000 seconds later still, D will have moved through the angle DGD′ = 37.5 degrees.

Appendix 1

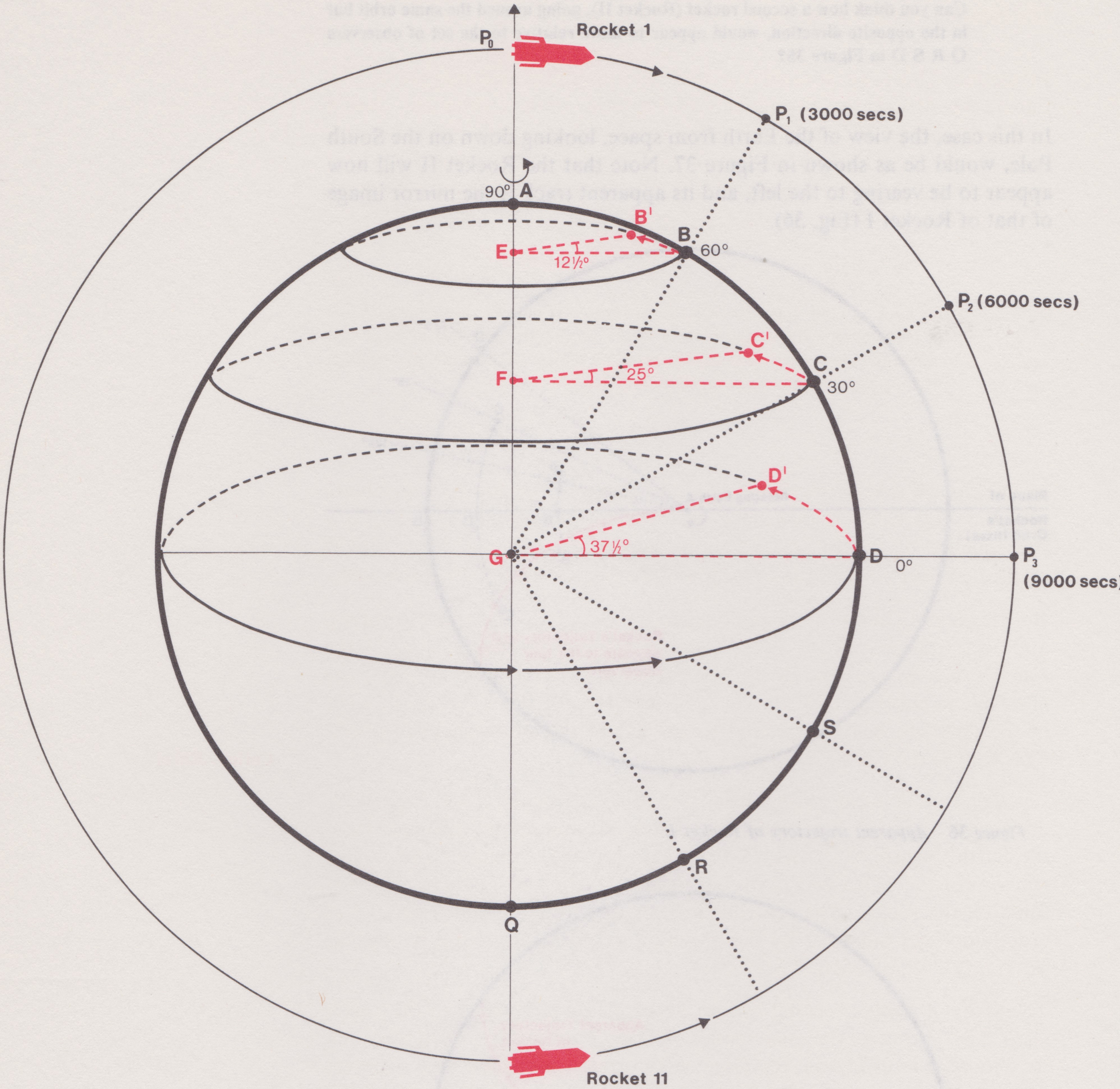


Figure 35 Illustration of the Coriolis Effect.

Viewed from above the North Pole, the successive positions of B, C and D, 3 000, 6 000 and 9 000 seconds after the rocket was at P_0 on the Earth's axis, above A, will be as shown in Figure 36.

How will things seem to our four observers? If they look at each other, they observe that they are all fixed firmly to the Greenwich meridian—they are stationary. But the rocket seems to them to be drifting westwards, in the opposite direction to the rotation of the Earth. To them, they are where they were to start off with, along the line ABCD, and it is the rocket that has drifted along the line A B″ C″ D″.

Can you think how a second rocket (Rocket II), going around the same orbit but in the opposite direction, would appear to move relative to the set of observers Q R S D in Figure 35?

In this case, the view of the Earth from space, looking down on the South Pole, would be as shown in Figure 37. Note that the Rocket II will now appear to be veering to the left, and its apparent track is the mirror image of that of Rocket I (Fig. 36).

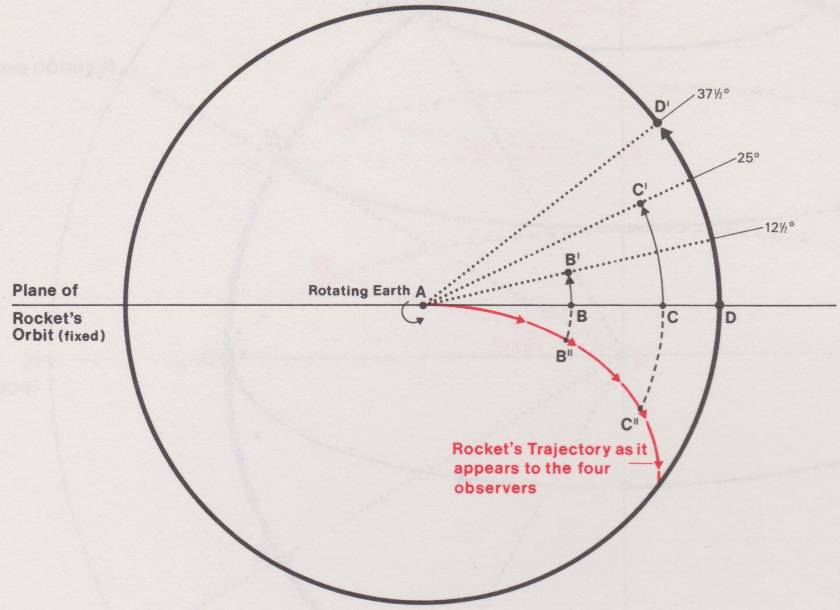


Figure 36 *Apparent trajectory of Rocket I.*

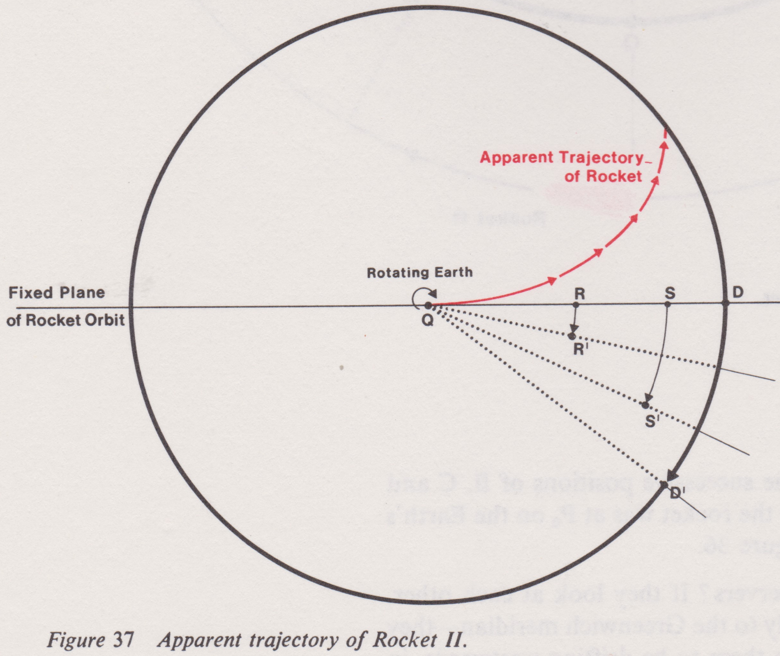


Figure 37 *Apparent trajectory of Rocket II.*

Perhaps the above explanation of the Coriolis effect has left you with a feeling that it is a rather abstract matter of how things *seem* to a set of

observers—a sort of optical illusion and not a real phenomenon at all. Actually it was real enough to make the first long-range artillery shells miss their targets by an embarrassingly large margin!

You can convince yourself of the reality of the effect by some quite simple experiments with more homely objects than shells and rockets. Put a plain circular piece of card or paper on your gramophone turntable and mark it with the three observers A, B and C along a radius at one-third and two-thirds of the distance from the centre to the edge, and at the edge. These will represent our observers at 60° N, 30° N and the equator respectively. Now spin the turntable slowly in an anti-clockwise direction and attempt to draw a line from the centre pin outwards through the three observer points. You will find that this is virutally impossible to do and that in every case you will get a line which is curved to the right away from the observers.

Now spin the turntable clockwise and see what happens.

If you are still not convinced and in need of a little exercise then go with a friend to the local park and sit in the middle of a roundabout. Get him to sit at the edge of the roundabout. While the roundabout is moving anti-clockwise, attempt to throw a ball to him. You will find it very difficult to do so. The ball thrown straight at the person on the margin of the roundabout will, as with our rocket, veer off to the right. You will, of course, after a short lapse of time become used to this and be able to 'lay-off' by throwing the ball initially somewhat to the left of your catcher and it will, taking the Coriolis effect into account, come back to the right place and hit the target.

A gas molecule in the atmosphere is in a somewhat similar situation to the orbiting rocket, inasmuch as it is attracted only towards the centre of the Earth, not towards any particular point on the Earth's surface. So air masses moving in the northern hemisphere tend to rotate in a clockwise sense, viewed from above, and in the southern hemisphere they tend to rotate counterclockwise. This can be seen very clearly in photographs of cyclones taken from satellites.

Appendix 2 (Black)

The velocity of waves

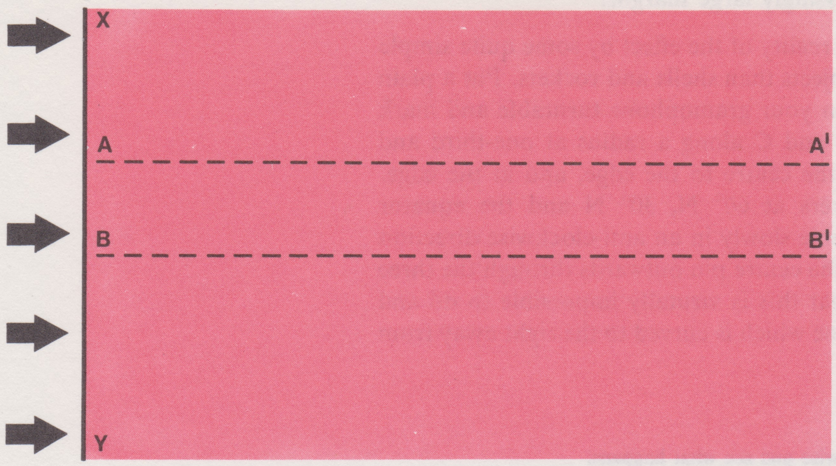


Figure 38 Longitudinal waves in a solid.

(*a*) *Longitudinal waves*

In Figure 38, XY is the left-hand face of a large solid slab. Consider in particular a section of the solid bounded by the planes AA′ and BB′. At a given instant of time ($t = 0$) we apply a uniform force to the XY face in the direction shown. Like any other section of the solid, that between AA′ and BB′ will begin to suffer a contraction of its longitudinal dimensions (left to right in the diagram). It will also try to expand its lateral dimensions. However, the planes AA′ and BB′ are also acted upon by forces exerted by the sections of the solid immediately above and below our section. These sections are also trying to expand their lateral dimensions. The forces on the two planes are equal and opposite so the net effect is that the planes do not move; all movement is confined to the longitudinal direction.

Suppose at the time $t = 0$ the face XY is set moving with a constant velocity, v. The applied force partially compresses the first vertical layer of the solid medium and sets it moving with the same velocity. This in turn pushes against the second layer, compresses it, and sets it moving, and so on.

The disturbance, or wave, travels on ahead of the XY face with a velocity u. Thus at time t the situation is shown in Figure 39 where XY has moved a distance v and the disturbance a distance u. All the material in the section between the planes XY and PQ is now moving with a velocity v.*

* *If you find it difficult to see why* two *velocities are involved,* u *as well as* v, *the following analogy might help. Think of a line of railway trucks. An engine moving with a velocity* v *comes up and starts shunting them. Although each truck in turn is set moving only with speed* v, *the disturbance caused by the engine is rapidly transmitted (with velocity* u) *to the other trucks. In other words a truck at a position well in advance of the engine will receive a jolt* before *the engine arrives at that position. The jolt therefore moves on ahead of the engine with a velocity* u, *which is greater than the velocity of the engine.*

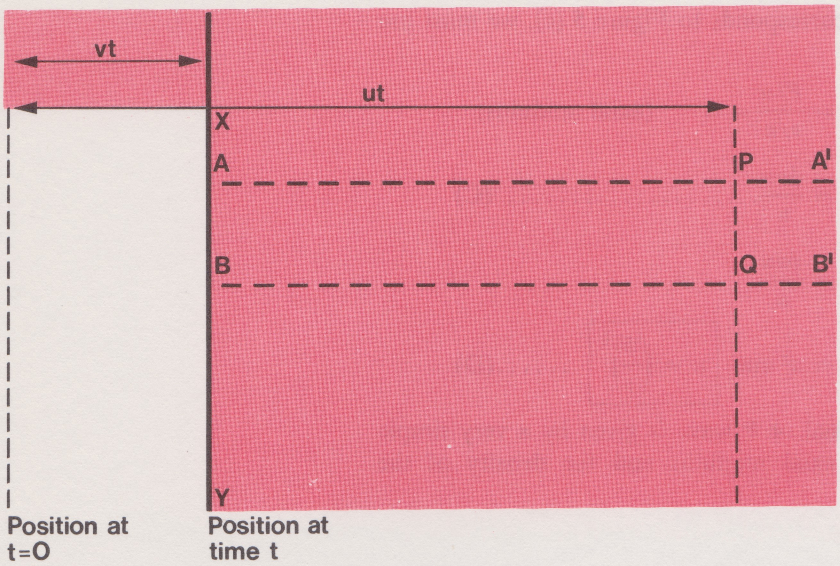


Figure 39 *Velocity of longitudinal waves.*

Why don't particles of the medium between XY and PQ oscillate backwards and forwards about their mean positions as happened in the case of sound waves generated by a vibrating metal strip in Unit 2?

This is a different situation to the one encountered in Unit 2, where we considered longitudinal sound waves produced by a vibrating metal strip. There the source of the disturbance moved forwards and backwards and this caused the particles of the medium also to move forward and backward—that is, to vibrate about their mean positions. Here the source of the disturbance is moving in one direction only and at a steady velocity; the particles of the medium, therefore, will do the same.

We can work out the value of the velocity u in the following way. Let ρ be the density of the uncompressed solid, and let α be the cross-sectional area presented by the section between AA′ and BB′ to the force. Then the mass of the moving part of the section is the original density, ρ, multiplied by the original volume, αut—that is, $\rho\alpha ut$.

The longitudinal momentum is given by the mass times its velocity:

$$\text{momentum} = \rho\alpha utv \quad \ldots\ldots\ldots\ldots\ldots\ldots\text{(A)}$$

This momentum was imparted by the force, F, applied to the section lying between AA′ and BB′. This force can be related to the momentum by the definition given in Unit 4. There, force was defined as the rate of change of momentum with time. In the present problem the momentum has changed from an initial value of zero to the one given by equation (A) and this occurred in a time, t.

Therefore, $$F = \frac{\text{momentum}}{\text{time}} = \frac{\rho\alpha utv}{t} = \rho\alpha uv \ldots\ldots\text{(B)}$$

Although this is an equation involving the quantity we wish to evaluate—that is, u—it is not a very helpful one. To get rid of quantities like F and α we must look for a second equation involving them. This is provided by the fact that the force F acting over the area α not only sets the medium moving but also compresses it. The stress producing the compression is given by

$$\text{stress} = \frac{F}{\alpha}$$

$$\text{and the strain} = \frac{\text{contraction}}{\text{original length}} = \frac{vt}{ut} = \frac{v}{u}$$

As the case we are considering corresponds to Figure 5 (c), we must use the axial modulus (ψ):

$$\frac{\text{stress}}{\text{strain}} = \frac{F/\alpha}{v/u} = \psi \qquad \text{(axial modulus)}$$

That is: $$F = \frac{\psi v \alpha}{u} \ldots\ldots\ldots\ldots\ldots\ldots\ldots \text{(C)}$$

So combining (B) and (C): $$\rho \alpha u v = \frac{\psi v \alpha}{u}$$

So $$\frac{\psi v \alpha}{\rho v \alpha} = u^2 \text{ and } \boxed{u = \left(\frac{\psi}{\rho}\right)^{\frac{1}{2}}} \quad \ldots\ldots\text{(D)}$$

Thus the velocity of a longitudinal or P wave is given by a very simple expression involving only the axial modulus and the density of the medium.

(*b*) *Shear waves*

A similar kind of argument can lead us to an expression for the velocity of a shear wave. In Figure 40 (a), a long strip of the medium with cross-sectional area α is initially at rest. At time, $t = 0$ a steady shearing force, F, is exerted across the end face, X, and this causes the end face to move upwards with velocity, v. This movement produces a shear strain in the first segment of the strip and sets it moving upwards with the same velocity. This segment in turn acts on the second, and so on along the strip (see Fig. 40 (b)). The velocity of the disturbance passing along the strip is w.

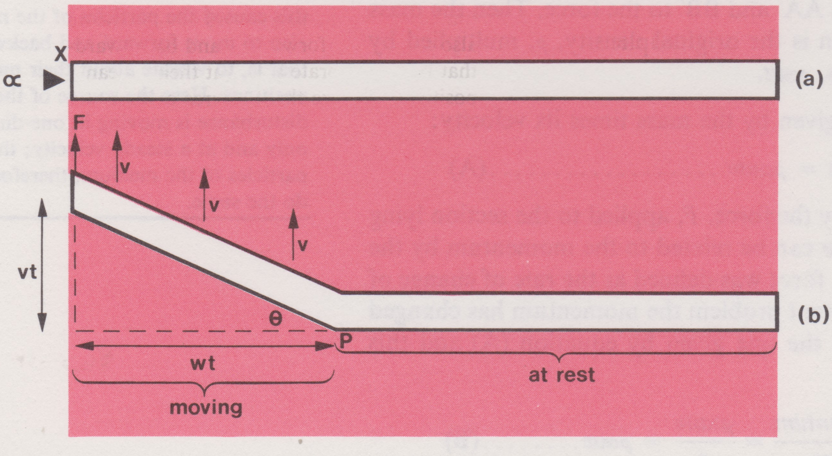


Figure 40 *Velocity of shear waves.*

After time, t, the mass of the moving part of the medium is: $\rho \alpha w t$, and the upward momentum is:

$$\text{momentum} = \rho \alpha w t v \ldots\ldots\ldots\ldots\ldots\ldots\ldots \text{(A}'\text{)}$$

One again the force is given by the rate of change of momentum.

Therefore, $$F = \frac{\rho \alpha w t v}{t} = \rho \alpha w v$$

Can you give a second equation involving F?

As before, we can obtain a second expression for F from studying the deformation of the medium.

What is the strain in this case?

The strain produced in the medium between X and P is given by the angular deformation θ, where

$$\theta = \frac{vt}{wt} = \frac{v}{w}$$

The stress is F/α as before.

Therefore, from equation (A′): $\dfrac{\text{stress}}{\text{strain}} = \dfrac{F/\alpha}{v/w} = \mu$ (the rigidity modulus)

Therefore, $$F = \frac{\mu y \alpha}{w} \quad \ldots\ldots\ldots\ldots\ldots(\text{B}')$$

From equations (B) and (B′) derive an expression for w.

$$\frac{\mu v \alpha}{w} = \rho \alpha w v$$

So, $$\frac{\mu v \alpha}{\rho \alpha v} = w^2$$

Therefore, $$\boxed{w = \left(\frac{\mu}{\rho}\right)^{\frac{1}{2}}} \quad \ldots\ldots\ldots\ldots(\text{C}')$$

Appendix 3 (White)

Glossary

BASALT A fine-grained, dark-coloured igneous rock. An igneous rock is formed by solidification from a molten or partially molten state.

ECLIPTIC The plane containing the Earth's orbit and the Sun.

ECLOGITE A granular rock composed essentially of garnet* (about 50 per cent) and pyroxene* (about 50 per cent).

GABBRO A coarse-grained, dark-coloured igneous rock.

GRANITE A coarse-grained, light-coloured igneous rock.

PERIDOTITE A rock composed essentially of olivine* (about 85 per cent) and pyroxene* (about 15 per cent).

* *For further details of rock-forming minerals, see* Understanding the Earth, *Chapter* 1, *'Minerals and Rocks' by K. Cox.*

Self-Assessment Questions

Section 22.2

Question 1 (*Objective 1*)

Which of the following statements are true?

(a) The outer planets are denser than the inner planets.
(b) The inner planets are denser than the outer planets.
(c) The outer planets have atmospheres dominated by hydrogen and its compounds.
(d) As a general rule the terrestrial planets have extremely rapid rotation rates.
(e) The outer planets have extremely rapid rotation rates.

Section 22.3.2

Question 2 (*Objective 1*)

Why does the Earth 'bulge' at the equator?

(a) Because of the 'pull' exerted by the Moon.
(b) Because of the centrifugal force exerted by the Earth's own rotation.
(c) Because of the law of gravitation.
(d) For none of the above reasons.

Section 22.4.2

Question 3 (*Objective 2*)

The average density of the Earth is 5.5×10^3 kg m^{-3}. Which of the following statements concerning the relative density of the Earth's interior are true?

(a) The density of Earth's interior is greater than that of the surface matter of the Earth.
(b) The densities of the Earth's interior and surface are about the same.
(c) The density of the Earth's surface is greater than that of the Earth's interior.

Section 22.4.3

Question 4 (*Objective 3*)

What useful data about the Earth's interior can be deduced from the study of earthquake waves?

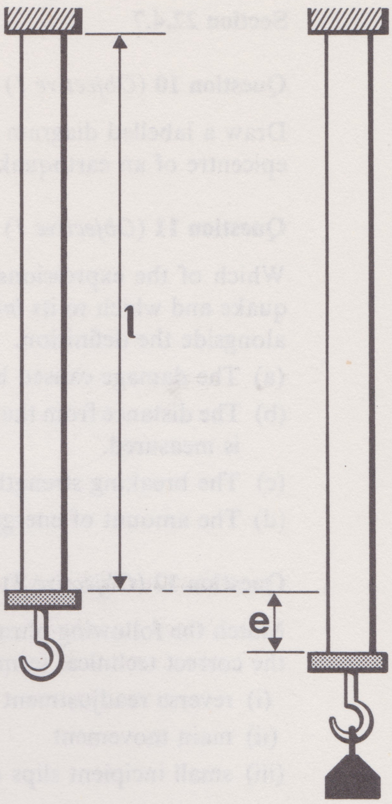
(a) density
(b) compression
(c) rigidity
(d) compressibility
(e) temperature
(f) pressure

Self-Assessment Questions

Section 22.4.4

Question 5 (*Objective 1*)

The diagrams here show the change in length of a piece of wire clamped at one end when a weight is attached to the wire. Which of the formations below defines the *strain* applied to the wire by the weight?



(a) e

(b) $l+e$

(c) e/l

(d) l/e

Question 6 (*Objective 4*)

A steel shaft 3.5 m long and 0.2 m in diameter is part of a hydraulic jack used for raising cars in a garage. When supporting a car weighing 1 450 kg the length of the shaft decreases by 7.65 μm. What is Young's modulus for the steel? Express your answer in terms of one of the units listed below.

(a) N; (b) N m^{-1}; (c) N m^{-2}; (d) N m^{-3}; (e) N m; (f) N m^2; (g) N m^3.

Question 7 (*Objective 4*)

Two wires, A and B, of the same diameter are joined together end to end. The material of wire A has a Young's modulus of 20.7×10^{10} N m^{-2}, and that of wire B is 12.3×10^{10} N m^{-2}. When the combination is subjected to a tensile force of 1 N, the total extension is found to be 0.65 mm. If the force is increased to 1.4 N, what is the accumulated extension of wire B, expressed in millimetres?

Either (i) record your answer;

or (ii) if unable to reach an answer choose one of the following reasons.

(a) Hooke's Law is not obeyed for a combination of materials.

(b) The diameter of the wires has not been specified.

(c) The length of wire B has not been specified.

(d) The ratio of the lengths of the two wires has not been specified.

Section 22.4.5

Question 8 (*Objective 1*)

Draw a diagram to show the directions of shear waves and compressional waves when the Earth fractures as the result of an earthquake.

Section 22.4.6

Question 9 (*Objective 1*)

The *epicentre* of an earthquake is defined as:

(a) the point of origin of the earthquake;

(b) a point on the Earth's surface vertically above the focus;

(c) the line joining the focus to the Earth's surface;

(d) the area where most damage is caused.

Section 22.4.7

Question 10 (*Objective 1*)

Draw a labelled diagram to show the relationship between the focus and epicentre of an earthquake, and the isoseismal lines produced.

Question 11 (*Objective 1*)

Which of the expressions below refers to the *magnitude* (1) of an earthquake and which to its *intensity* (2). Answer by putting the number 1 or 2 alongside the definition.

(a) The damage caused by the earthquake.

(b) The distance from the epicentre to the point where intensity/magnitude is measured.

(c) The breaking strength of the rocks in which the earthquake occurs.

(d) The amount of energy released by the earthquake.

Question 12 (*Objective 1*)

Match the following phrases concerning earthquakes, listed on the left, to the correct technical terms listed on the right.

(i) reverse readjustment movement	(a) compression shock
(ii) main movement	(b) aftershock
(iii) small incipient slips or fracture	(c) foreshock
	(d) shear shock
	(e) principal shock

Section 22.4.8

Question 13 (*Objective 1*)

Why are three seismometers usually used to pick up earthquake waves?

(a) To record the amplitude and direction of both the P and S waves.

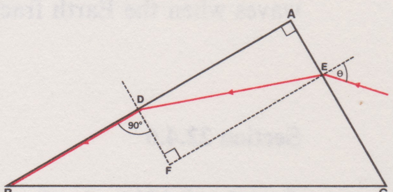
(b) As a fail-safe device in case of instrument defect.

(c) So that minor differences in the amplitude and direction of the P and S waves within the recording station can be detected.

Section 22.4.11

Question 14 (*Objective 4*)

A ray of light enters a right-angled prism through the face AC (see Figure) at an angle of incidence of θ and emerges from the face AB at an angle 90° to the normal. Derive a relation between θ and the critical angle, ϕ, for the glass of the prism.



Indicate your answer from among the following possibilities:

(a) $\sin\theta = \cos\phi$

(b) $\sin\theta = \cot\phi$

(c) $\sin\theta = \tan\phi$

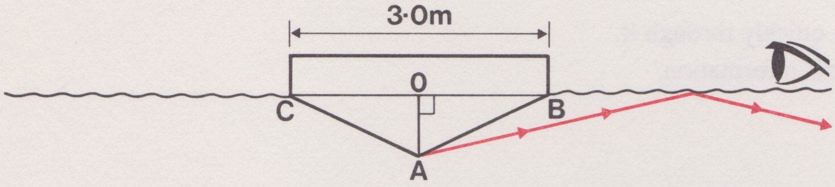
(d) $\cos\theta = \tan\phi$

(e) $\cos\theta = \cot\phi$

(f) $\cos\theta = \sin\phi$

Question 15 (*Objective 4*)

The diagram shows a section through a boat of width 3.0 m. When the boat floats on water, its underneath sides, AB and AC, cannot be seen by a viewer at the side no matter where he positions his eye above the water level. What does this allow you to say about the length AB (or AC)? The refractive index of water is 1.333.



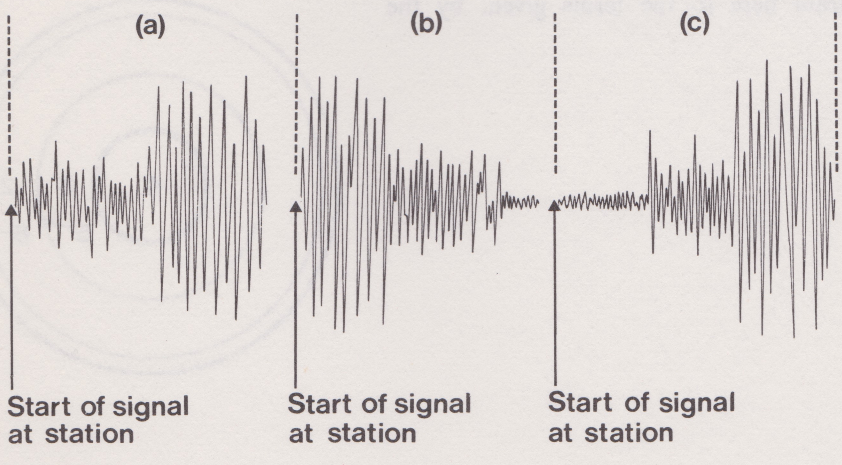
Question 16 (*Objective 4*)

Using the data provided in the Table 2 of the Unit, find the critical angle for P waves at the boundary between the core and the mantle.

Section 22.5.1

Question 17 (*Objective 1*)

Which of the three tracings shown below represents a seismogram of an earthquake recorded at about 10 000 km from the epicentre?



Section 22.5.5

Question 18 (*Objective 7*)

Why do S waves not pass through the Earth's core?

(a) They are all reflected from its surface.

(b) The outer part of the core is liquid.

(c) They turn into P waves.

(d) None reach this depth.

Section 22.5.6

Question 19 (*Objective 7*)

Why is the centre of the Earth considered to be solid?

(a) Because pressures are so great that any liquid would be squeezed into a solid.

(b) Because S waves pass through it.

(c) On meteorite evidence.

(d) Because P waves travel more quickly through it.

(e) It has been proved by borehole information.

Section 22.5.7

Question 20 (*Objective 1*)

On what evidence was the existence of the Mohorovičić discontinuity first suggested?

(a) On the reflection of P waves from this surface.

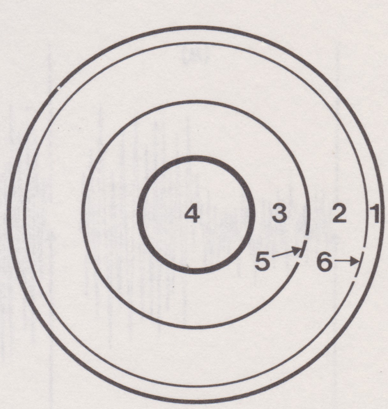
(b) On the recording of two distinct sets of P and S wave deflections at recording station near the earthquake.

(c) On the fact that S waves will not pass through this discontinuity.

(d) Evidence provided by the 'Mohole' project.

Question 21 (*Objective 5*)

Match the number in the diagram here to the terms given, by the appropriate number.



(a) mantle

(b) Mohole

(c) Mohorovičić discontinuity

(d) inner core

(e) core-mantle boundary

(f) outer core

(g) crust

Section 22.5.8

Question 22 (*Objective 1*)

Match the list of regions given on the left to the correct depth to the Moho from the list on the right.

(i) Under ocean basins (from the sea-floor).	(a) 5 km.
(ii) Under continental mountain ranges.	(b) 30–35 km.
(iii) Under extensive flat areas of continent.	(c) 8–10 km.
	(d) 60–65 km.
	(e) 100 km.

Question 23 (*Objective 1*)

The lower part of the Earth's continental crust is probably:

(a) granite

(b) basalt

(c) peridotite

(d) unknown

Section 22.5.9

Question 24 (*Objective 1*)

Select from the list below the reasons why the upper mantle is thought to be peridotite.

(a) It has the right rigidity, compressibility, and density.

(b) It forms the floor of the deep ocean basins.

(c) Fragments of peridotite are brought to the surface by volcanoes.

(d) It is the only natural rock that could occur at the temperature and pressures in the upper mantle.

(e) Volcanic islands are formed of peridotite.

(f) There are meteorites of peridotite composition.

Section 22.5.10

Question 25 (*Objective 1*)

One of the currently favoured explanations for the presence of the low velocity layer is:

(a) That the intersection of the peridotite melting point curve with the thermal gradient causes partial melting.

(b) A liquid layer occurs at that depth.

(c) There is a layer of eclogite at this depth.

(d) That the peridotite at this depth is perfectly plastic.

Question 1

Answers (b), (c) and (e) are true.

Comment These are straightforward statements of fact as given in section 22.2.

Question 2

Answer (b)

Comment The major force which is causing the Earth to bulge at the equator is the centrifugal force exerted by the Earth's own rotation. It should also be realized, however, that the pull, the gravitational pull, exerted by the Moon is trying to pull the equatorial plane, the plane at right angles to the Earth's rotational axis, into line with the ecliptic. These two forces must be identified in your own mind as being completely separate.

Question 3

Answer (a)

Comment Here we give you the average density of the Earth and ask you which of the statements concerning the relative density of the Earth's interior is true. This is surely very straightforward indeed. The density of the Earth's interior has got to be greater than the density of the surface material for the density of the surface material is only 2.9×10^3 kg m^{-3}.

Question 4

The information that we can get about the Earth's interior from a study of earthquake waves basically concerns the rigidity and the compressibility of the material that forms the Earth's interior. We cannot get any information at all directly concerning the temperature, the pressure, or the density. We have to know these factors before we can determine the rigidity and the compressibility. You should therefore have marked rigidity and compressibility. However, unless you read the text very carefully, you could very easily have marked item (a) and if you have done so, then credit yourself with a half mark for being on the right track.

Question 5

The answer to this question is simply the definition of *strain*, which is:

$$\text{strain} = \frac{\text{extension}}{\text{original length}} = \frac{e}{l}$$

The correct answer is thus (c).

Question 6

Answer 2.07×10^{11} N m^{-2}.

Comment Because of the form of the units, the force must be expressed in newtons, that is, the mass of the car in kilograms must be multiplied by the acceleration due to gravity (9.81 m s^{-2}).
So, the force, $F = 1450 \times 9.81 = 1.422 \times 10^4$ N.

The expression for Young's Modulus, Y, in terms of the area under stress, A, the length of the stressed object, l, and the change in length, Δl, is

$$Y = \frac{E/A}{\Delta l/l} = \frac{El}{A\,\Delta l}$$

If the diameter of the shaft is 0.2 m, then the area,

$$A = \frac{\pi \times (0.2)^2}{4} = 0.0314 \text{ m}^2.$$

l is given as 3.5 m, and Δl is 7.65×10^{-6} m

So,

$$Y = \frac{(1.422 \times 10^4) \times 3.5}{0.0314 \times 7.65 \times 10^{-6}} = \frac{4.98 \times 10^4}{0.240 \times 10^{-6}} = 2.07 \times 10^{11} \text{ N m}^{-2}$$

Question 7

Answer (ii) (d)

Comment Hooke's Law *is* obeyed for the combination of wires but there is insufficient information to arrive at the extension of wire B. Specifying the diameter of the wires or alternatively the length of wire B would still not have made it possible to solve the problem. However, if the ratio, r, of the length of wire A to the length of wire B had been specified then a solution would have been possible, as can be seen in the following way:

$$Y = \frac{Fl}{A\Delta l}$$

For wire A, of length l_a, under a tensile force, F, of 1 N:

$$20.7 \times 10^{10} = \frac{1 \times l_a}{A \times \Delta l_a} \quad \ldots\ldots\ldots\ldots\ldots(1)$$

For wire B, of length l_b, under the same stress:

$$12.3 \times 10^{10} = \frac{1 \times l_b}{A \times \Delta l_b} \quad \ldots\ldots\ldots\ldots\ldots(2)$$

Dividing (2) by (1) we have:

$$\frac{12.3 \times 10^{10}}{20.7 \times 10^{10}} = \frac{1 \times l_b}{A \times \Delta l_b} \times \frac{A \times \Delta l_a}{1 \times l_a}$$

$$\frac{12.3}{20.7} = \frac{1}{r}\frac{\Delta l_a}{\Delta l_b}$$

But the total extension $(\Delta l_a + \Delta l_b) = 0.65 \times 10^{-3}$ m

$$\therefore \Delta l_a = 0.65 \times 10^{-3} - \Delta l_b$$

$$\frac{12.3}{20.7} = \frac{1}{r}\frac{(0.65 \times 10^{-3} - \Delta l_b)}{\Delta l_b}$$

If r had been given, Δl_b could have been obtained from this relation. Under a tensile stress of 1.4 N, the extension of wire B would have been $1.4\ \Delta l_b$. This is the quantity you were asked to calculate—but you had not been given the value of r!

Question 8

The diagram you should have here is Figure 9 in the text.

Question 9

Answer (b)

Comment An epicentre is a point on the surface of the Earth vertically above an earthquake focus. It is a theoretical point which can sometimes be deduced by actual observations on the surface of the Earth; and in other cases it has to be determined by observations from seismic stations.

Question 10

The diagram you should have here is Figure 11 in the main text.

Question 11

This is just a matter of having read the text correctly. The earthquake's *intensity* is measured on an arbitrary scale indicated by the damage that is done by the earthquake, whereas the *magnitude* is controlled by the amount of stress that builds up within the Earth before the Earth fractures. This, of course, relates directly to the breaking strength of the rocks. The greater the breaking strength in the rock the more stresses will be allowed to build up before that rock fractures and the greater will be the magnitude of the resulting earthquake. If the rocks in which the stresses are being accumulated is weak, such as a clay or a gravel, then the stresses will be of no great magnitude before the Earth yields, and an earthquake is caused. In this case the magnitude would be low. You should therefore have indicated 1 against item (c) and 2 against item (a).

Question 12

Here you were required to match phrases cast in informal jargon against technical terms. The first of these was a reverse readjustment movement. Very commonly when an earthquake happens the actual amount of movement along a fault is greater than that needed to accurately compensate for the stress. What happens is that the movement overshoots the neutral point and therefore small readjustment movements have got to take place. These, of course, occur after the main movements and reverse readjustment movements are referred to as *aftershocks.* You should therefore have put against phrase (i) the letter (b). Main movement—this of course refers to the major movement along a fault and this causes the principal shock. You should therefore have put against item (ii) the letter (e). Small incipient slips or fractures are the fractures that are caused before the major movement and are referred to as *foreshocks.* You should have inserted against item (iii) the letter (c).

Question 13

Answer (a)

Comment Three seismometers are used to pick up earthquake waves because two phenomena have to be recorded. A vertical seismometer is required to detect the time of arrival of the vertical vibrations of the Earth, as the P waves come into the station vertically from below. You will recall that the S waves are transverse waves oscillating in a direction at right angles to

the direction of their propogation. We require to know in which direction they are oscillating. Therefore, so far as the S waves are concerned, we not only need to know the time of their arrival but also the direction of their oscillation. This can only be obtained by having two horizontal seismometers situated so that they are at right angles to each other.

Question 14

Answer (b)

Comment The ray of light strikes the face AB at the critical angle (because it emerges at an angle 90° to the normal). Therefore angle EDF = ϕ.

The angles of the triangle DEF add up to 180°.

$\therefore$ angle EDF + angle DFE + angle FED = 180°

$\therefore$ ϕ + 90° + angle FED = 180°

$\therefore$ angle FED = 90° − ϕ.................. (1)

Applying Snell's Law of Refraction at the point where the ray enters the face AC, we have:

$$\frac{\sin\theta}{\sin(\text{angle FED})} = n \;.......\; (2)$$

Substituting (1) in (2), we have:

$$\frac{\sin\theta}{\sin(90-\phi)} = n$$

that is,

$$\frac{\sin\theta}{\cos\phi} = n \quad ....(3)$$

But the refractive index, n, and the critical angle, ϕ, are related by

$$\frac{1}{\sin\phi} = n \quad ....(4)$$

Substituting (4) in (3):

$$\frac{\sin\theta}{\cos\phi} = \frac{1}{\sin\phi}$$

that is, $\sin\theta = \cot\phi$

Question 15

Answer AB (or AC) is shorter than 2.0 m.

Comment The fact that the underfaces cannot be seen means that the point A cannot be seen. This is presumably because any ray of light from A that travels to the air-water interface is internally reflected and so remains in the water (see Figure).

The ray that makes the largest angle of incidence to the air-water interface is that which skims along the line AB. We can therefore conclude that the line AB meets the water surface at

an angle at least as great as the critical angle, i_c (otherwise the ray close to AB will emerge into the air and be observed).

Therefore, angle OBA $\leqslant 90 - i_c$ ....(1)

$$\cos \text{OBA} = \frac{\text{OB}}{\text{AB}} = \frac{1.5}{\text{AB}} \quad \ldots\ldots\ldots\ldots\ldots\ldots(2)$$

Because of equation (1) the cosine of angle OBA must be greater than the cosine of $(90-i_c)$. (Note that between 0° and 90°, the cosine of an angle *decreases* as the angle increases.)

$$\therefore \cos(90-i_c) < \frac{1.5}{\text{AB}}$$

$$\sin i_c < \frac{1.5}{\text{AB}}$$

$$\text{AB} < \frac{1.5}{\sin i_c}$$

Because the sine of the critical angle is equal to the inverse of the refractive index.

$$\text{AB} < 1.5 \times 1.33$$
$$\text{AB} < 2.0 \text{ m.}$$

Question 16

Answer 36.5°

Comment From equation (13) in the text, the sine of the critical angle, i_c, is given by the ratio of the velocities of the waves on either side of the boundary, that is:

$$\sin i_c = \frac{V_i}{V_r}$$

The velocities of the P waves in the outer core and lower mantle are given in Table 2 as 8.1 and 13.6 respectively.

$$\sin i_c = \frac{8.1}{13.6} = 0.595$$

$$i_c = 36.5°$$

Question 17

Of the three tracings roughly represented here, the first shows both P and S waves, the second shows a single series of waves and the third shows P and S waves. As the epicentre of the earthquake was 10 000 km from the recording station it is obvious that (c) is the correct trace because it has the longest 'lead in' before the start of the P waves and an interval before the S waves.

Question 18

Answer (b)

Comment S waves are, of course, transverse waves which do not transmit through fluids.

Question 19

Although the evidence put forward is by no means definite the major evidence we have is that P waves travel more quickly through the very centre of the Earth than they do through the outer part of the Earth's core. So, as we know that P waves travel more quickly through a solid than they do through a fluid, we assume that the centre part of the Earth is a solid. A second consideration is that the pressures are so great within the centre of the Earth that whatever the composition of the Earth's core, be it metallic, silicate, or hydrogen and helium, these would be compressed into a solid under pressures obtaining there. You should therefore have (b) as a first choice and (a) as your second choice.

Question 20

Answer (b)

Question 21

Answers	(a) mantle	(2)
	(c) Mohorovičić discontinuity	(6)
	(d) inner core	(4)
	(e) core-mantle boundary	(5)
	(f) outer core	(3)
	(g) crust	(1)

The only item not listed above is (b) Mohole. This is, of course, not a part of the Earth at all but was the United States project (now abandoned) to drill a hole down to the Earth's mantle.

Question 22

Answers (i) matches with (a)
(ii) matches with (d)
(iii) matches with (b)

Question 23

We really know very little about the lower part of the Earth's continental crust, so the correct answer here is (d). You may think that basalt might be the correct answer and indeed you might be right—it could be basalt! It certainly is not granite or peridotite so far as we know; but it could well be highly metamorphosed rocks which we see occasionally cropping out at the surface of the Earth. The answer we expect here is 'unknown'. If you have in fact marked basalt then you are doing all right—many geologists would agree with you.

Question 24

You should indicate, first of all, that on the seismic evidence, peridotite has the right properties of rigidity, compressibility and density to fit for the deduced seismic velocities at that depth; and on these grounds alone peridotite is perfectly acceptable—so you should have (a). We also know that fragments of peridotite are brought to the surface by volcanoes. It is suggested that pieces of the Earth's mantle could very easily be torn from the sides of a volcanic conduit by the ascending magma, and be brought to the surface. This is indeed what we think has happened, so item (c) is further evidence. A third line of evidence, which is distinctly more tenuous,

is that there are meteorites of peridotite composition. So you should also mark item (f).

Question 25

Possibly the only reasonable explanation at the moment for the presence of the low velocity layer is that it represents the intersection of the peridotite melting point curve with the thermal gradient, thereby causing partial melting. The answer is thus (a). You should realize, however, that this is very much an idea rather than an hypothesis.

Acknowledgements

Grateful acknowledgement is made to the following sources for material used in these Units:

Unit 22

ILLUSTRATIONS

SEISMOLOGICAL SOCIETY OF AMERICA, California, U.S.A., from *Bull. Seismol. Soc. Am.*, 58, Fig. 10.

Unit 23

ILLUSTRATIONS

THE MACMILLAN COMPANY, New York, from *Handbook of Geophysics*, 1960, Figs. 8 and 9; U.S. DEPARTMENT OF COMMERCE, from *ESSA Technical Report IER 46-IES 1* for 'An Analysis of the Earth's Magnetic Field from 1835–1965' by K. L. McDonald and R. H. Gunst, Figs. 11, 14a and 14b.